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Ten Years Later...

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Just 10 years ago WOCE held its international Conference in Paris. Countries made statements there about the parts of the WOCE Implementation Plan that they intended to carry out. In the Conference report is a table comparing the Implementation Plan requirements with the commitments made. Now, with WOCE observations effectively ended we can see what we actually did!

The table below indicates that we met or exceeded our objectives and therefore that the commitments made by the 38 countries represented in Paris were real. However, the ultimate measure of the success of WOCE is not the number and quality of the observations but is the progress we make towards the WOCE objective of improving ocean models for climate research. Two articles in this Newsletter relate to this topic. Peter Saunders' shows that in the past 2 years half the articles referenced in the WOCE bibliography relate to modelling. Claus Böning reviews the WOCE/CLIVAR Workshop on Ocean Modelling for Climate held at NCAR this summer. Among its conclusions is a recommendation for the way forward on model intercomparisons, a topic about which we will hear much more in coming months.

The 25th WOCE SSG meeting

The WOCE SSG met in Brest in early October. It reviewed the progress towards the WOCE objectives in three areas – Data Assimilation, Air-Sea Fluxes and Ocean Modelling. The SSG was generally satisfied with the evidence of progress made. The SSG received reports on the Mercator and FOAM projects in which data assimilation was directed towards short term (few days) ocean forecasting rather than the ocean state estimation of interest to WOCE. In discussing air-sea fluxes and model development the SSG identified the need for WOCE-recommended forcing fields for the

period 1980–2007 to be developed and distributed. In the past 2 years WOCE deliberately slimmed-down its oversight structure of committees and working groups and in almost all regards this has been successful. However the Ocean Modelling Workshop highlighted the need for a group to ensure that ocean model development and validation received sufficient attention. In consequence a joint WOCE/CLIVAR Ocean Model Development Working Group is to be established.

The WOCE AIMS workshops appear to be stimulating the community to synthesise WOCE results and are fostering new collaborative research. Workshops on Variability and on Global Fluxes are now planned for the year 2000 as is the final WOCE Conference in 2003.

Thanks to John Church

WOCE-25 was the last at which John Church acted as WOCE co-chair. John assumed this role in 1994 and since then he has helped to steer WOCE through what could have been a difficult transition from observations to analysis and into the present strong position that WOCE AIMS holds in WCRP. I would like to personally thank John for this significant contribution which has been recognised at the WCRP level by his “promotion” to membership of the Joint Scientific Committee. The co-chairmanship of the SSG now passes to Bill Large and Peter Killworth.

Coming tasks needing your help

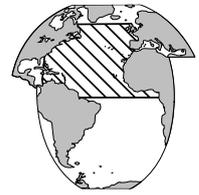
With observations effectively ended, the One-Time WHP data almost all submitted and the first CD-ROMs of WOCE data distributed, there is an urgent need to tidy up some loose ends. The first task is the compilation of a comprehensive inventory of all the observations made during WOCE. This will require detective work by the IPO, DIU and the DACs and SACs since it is now almost 10 years since the earliest observations were made and some memories may be fading and documentation hard to find. We need your co-operation in answering questions and digging in your records so that the inventory will be as complete and accurate as possible.

A related task is the compilation of the Repeat Hydrography data set. The WHPO and WHPPC rightly gave the One-Time data set priority but now the repeat observations, that will be essential for assessing temporal variability, must be collected and their quality be assessed. Again your co-operation will be required.

	<i>Implementation Plan needs</i>	<i>Conference commitments</i>	<i>Carried out</i>
One-time Hydro	10.4 ship years	21.5 ship years	14 ship years*
Repeat Hydro	15.0 ship years		15 ship years*
Subsurf. floats	ca. 3500	ca. 1900	2130
Surface drifters	ca. 4000	ca. 2000	4368
Moorings	ca. 350	ca. 250	330

*Based on number of stations reported.

Upper Ocean Temperature Structure in the Atlantic Ocean from Expendable Bathythermographs: WOCE and pre-WOCE, 1967–1996



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XBT data are provided by both research vessels and merchant ships, the latter operating as Voluntary Observing Ships (VOS). The observations serve both operational and research requirements. For example, NOAA uses the data in experimental climate forecasts (Ji et al., 1994), while WOCE investigators use the data to study Upper Ocean Thermal (UOT) variability (e.g., Molinari et al., 1997).

The Global Temperature Salinity Pilot Project (GTSP) was initiated by the Intergovernmental Oceanographic Commission and the World Meteorological Organisation to develop a protocol for providing the highest quality temperature data to forecasters and researchers. As part of the GTSP, three Data Assembly Centres (DAC) were created to quality control (QC) temperature profiles

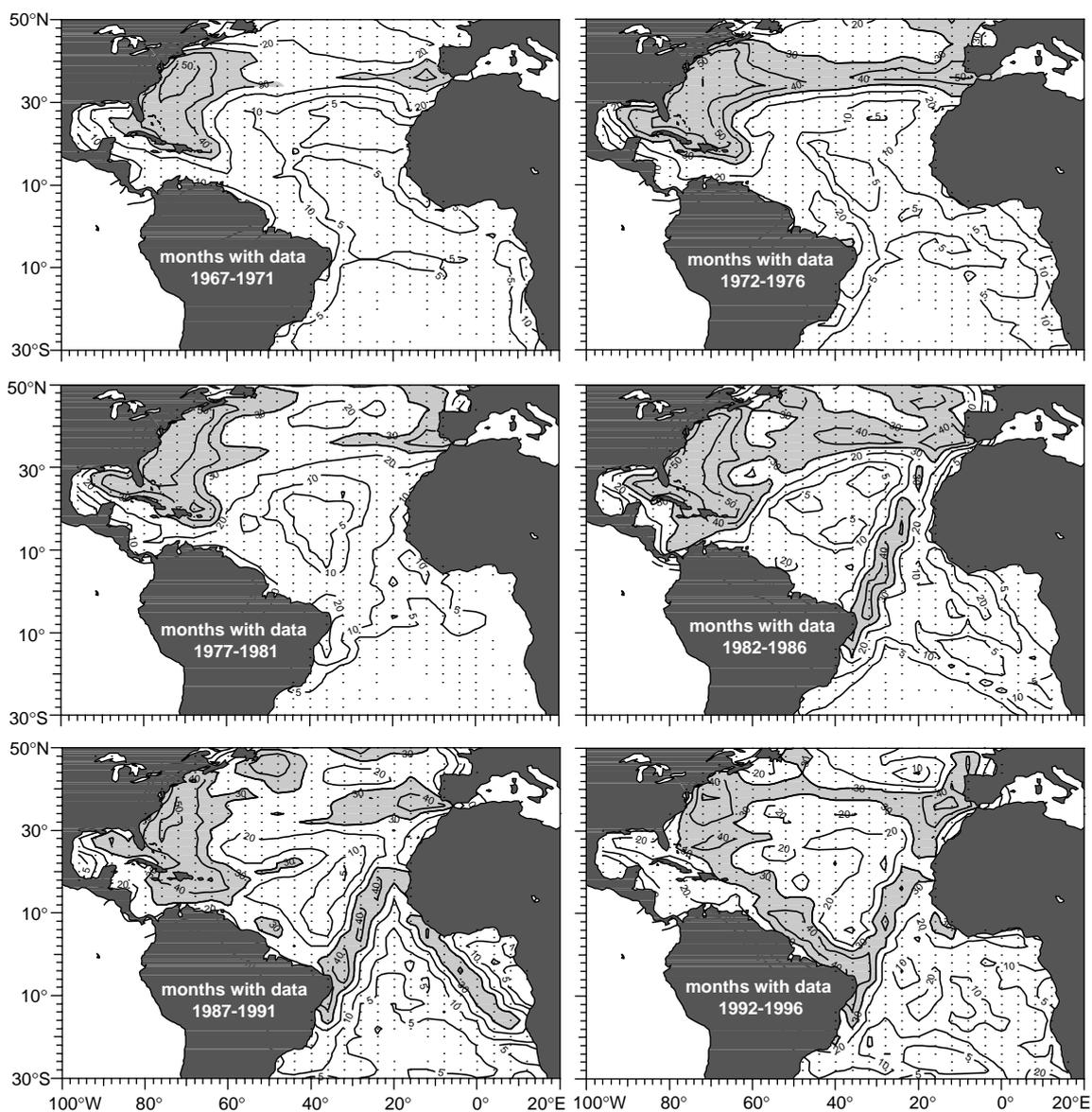


Figure 1. Data availability given on a 2° of latitude by 2° of longitude grid as months with data within five year bins (i.e., a maximum of sixty months). Grid points with data are marked by dots. A computer software package was used to contour the data distribution field, thus contours appear in regions with no data. The contour interval is uneven; 5, 10, 20, 30 data points, etc. Regions with greater than 20 data points are shaded.

from the global XBT network. NOAA's Atlantic Oceanographic and Meteorological Laboratory focused on the Atlantic Ocean, the Scripps Institution of Oceanography (SIO), on the Pacific Ocean and CSIRO, Hobart, Australia on the Indian Ocean.

Two types of XBT data are available. Real time data are typically low-resolution profiles (e.g., 29 depth-temperature pairs in a 450 m or 750 m cast) transmitted from ship-to-shore via satellite. Delayed mode data are typically higher resolution profiles (e.g., every 1 to 2 m) and represent the complete profiles forwarded to data collection centres after completion of a cruise. Initially, the DACs performed QC on both real time and delayed mode data. However, the QC of the real time data gradually was taken over by the Marine Environmental Data Service, MEDS, Canada and the DACs now concentrate on the delayed mode data.

Herein, we show summaries of the XBT data as well as products that were generated for the entire Atlantic to:

- (1) provide potential data users with information on the extent of available data throughout the Atlantic;
- (2) show climatological uses for the observations;
- (3) give a preliminary perspective of the WOCE UOT structure.

The WOCE-era quality controlled data are available from the WOCE data centre in Brest. The historical data are available from the US NODC.

Data and quality control

As described previously, XBT data were first collected in the Atlantic in 1966. However, extensive data collection, primarily in the north-western North Atlantic, did not begin until 1967. Early data collection was primarily confined to regional oceanographic surveys. More recently, as part of IGOSS, the use of VOS to collect UOT data became more prevalent. As part of WOCE, a series of transects were established in 1990 to provide data on the North Atlantic. The lines were based on a combination of scientific needs (e.g., resolving gyres) and available commercial tracklines. Optimally, each line was to be occupied monthly, with four probes launched per day along the transect. A consortium of US and international organisations have been collecting data along these lines since 1990. Unfortunately, for primarily logistical reasons many of the lines in the South Atlantic have not been occupied on a regular basis, as will become obvious.

The number of months with data in 5-year bins (60 months) on a 2° of latitude by 2° of longitude grid is shown in Fig. 1. The predominance of data availability in the mid-latitude western North Atlantic throughout the 1967–1996 period is one obvious feature of these distributions. Large data voids in the central subtropical North Atlantic and throughout the South Atlantic are also apparent. One success of WOCE was to begin additional data collection in these data sparse areas (i.e., compare the 1992–1996 data distribution in the South Atlantic with the 1987–1991 distribution in the same region).

To date the historical and WOCE-era data have not been subjected to a uniform quality control effort. WOCE real and delayed mode data (i.e., post-1990) in the Atlantic are quality controlled using GTSP procedures outlined in Daneshzadeh et al. (1994). "Best data-sets" that include both real and delayed mode profiles for 1990, 1991, 1992, 1993, 1994 and 1995 were quality controlled by the AOML DAC. These data thus include the quality control flags and meta-data specified by the GTSP (see Daneshzadeh et al., 1994 and 1996 for examples). The pre-1990, "historical data" were quality controlled prior to the introduction of the GTSP protocols and thus do not include much of the information available for the WOCE-era data. Real time data from 1996 through 1998 have been quality controlled by MEDS, with delayed mode data being compiled from these years for future review. To compensate for the lack of uniform quality control and to eliminate erroneous profiles missed in the QC of the pre-WOCE data, a standard deviation test was applied to all data. First the original profiles were interpolated to a standard profile with 50 m resolution. These data were then averaged by month and year onto a 2° of latitude by 2° of longitude grid. Record length monthly means and standard deviations were computed from these monthly means. Temperature values that were more than 2.5 standard deviations from the record length monthly means were flagged and not used in the analyses. A uniform quality control of both the historical and WOCE XBT data is planned.

Data products

Two types of data products are presented climatological UOT characteristics and UOT characteristics for the WOCE years, 1990–1996. All products are generated on a 2° of latitude by 2° of longitude grid with bi-monthly temporal resolution. Greater temporal resolution (i.e., monthly) for both synoptic and climatological conditions is possible in the Northern Hemisphere but not the Southern Hemisphere. January–February climatology of temperature at 200 m is given in Fig. 2. We wish to stress that although the fields are contoured to 40°S, data availability is quite limited south of 15°S to 20°S. The contours in these southern areas outside the data coverage must thus be viewed with some scepticism.

Mayer et al. (1998), hereinafter MMF, described the Atlantic gyre structure represented in the upper layer temperature fields and related the gyres to simple Sverdrup dynamics. In the South Atlantic, the main feature of the temperature fields is the subtropical gyre. As described by Peterson and Stramma (1991), for instance, this gyre extends from approximately 10°S to 40°S, depending on the depth. At 200 m, the gyre is bounded on the east by the temperature gradients associated with the Benguela Current. Approaching the equator, the gradients become associated with the South Equatorial Current (SEC). The SEC bifurcates at about 9°S, on the eastern tip of Brazil at 200 m. The northern branch of the SEC feeds the North Brazil Current and the southern branch, the Brazil Current. After separating from the coast to the south, the Brazil Current feeds the

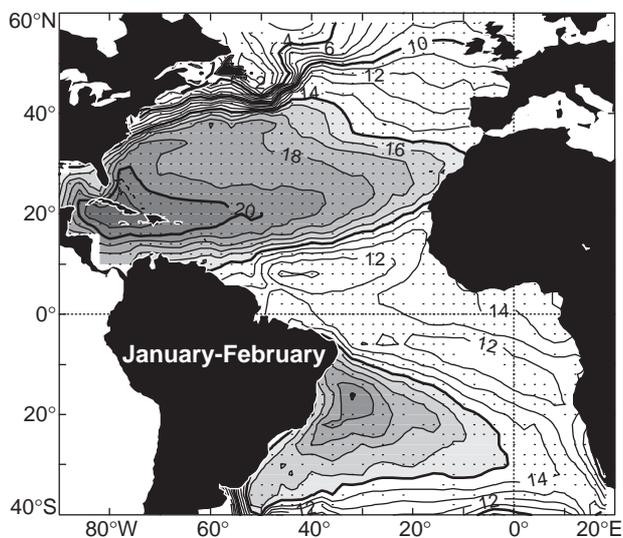


Figure 2. January–February mean temperature distribution at 200 m. Grid points with available data are marked by dots. The entire field is contoured by a computer software package, thus contours appear in data-void areas.

South Atlantic Current.

The structure of the subtropical gyre exhibits a vertical dependence similar to that observed in the vertical structure of the subtropical gyre of the North Atlantic (i.e., a polewards displacement of the gyre axis with depth, not shown). By 200 m, the axis has shifted southward to about 15°S to 20°S, which is somewhat to the north of the gyre position predicted by Sverdrup dynamics (MMF).

The equatorial gyre is located to the north of the subtropical gyre. At 200 m, this gyre extends approximately zonally across the western and central basin between 5°S and 10°S and then trends south-eastward into the boundary, east of the Greenwich meridian (Fig. 2). The gradients on the northern side of the equatorial gyre are associated with the South Equatorial Undercurrent. Sverdrup dynamics implies a simulated equatorial gyre centred somewhat to the north of the observed gyre at 200 m.

The tropical gyre is apparent just north of the equator. At 200 m, the temperature gradients on the southern side of the tropical gyre are associated with the North Equatorial Undercurrent (NEUC). In these temperature distributions, the NEUC is a permanent feature of the tropical circulation. The position of the tropical gyre implied from Sverdrup dynamics is close to the position inferred from the temperature distribution at 200 m (MMF). Finally, the observed North Atlantic subtropical gyre at 200 m is located equatorward of the position given by Sverdrup dynamics (MMF).

Annual harmonics have been computed from the temperature distributions for those grid points with more than 3 bi-months of data. The percent variance accounted for by the annual harmonic and the phase and amplitude of this component of variability are shown in Fig. 3 for 200 m. Except for a few isolated areas, the annual harmonic does not contribute significantly to variability below 200 m (not shown). Although, the percent variance fields are contoured for the entire region north of 40°S, data points indicate that data are too sparse south of 20°S to compute meaningful statistics for most of the region. At 200 m, maximum variance accounted for is typically about 60%, as compared

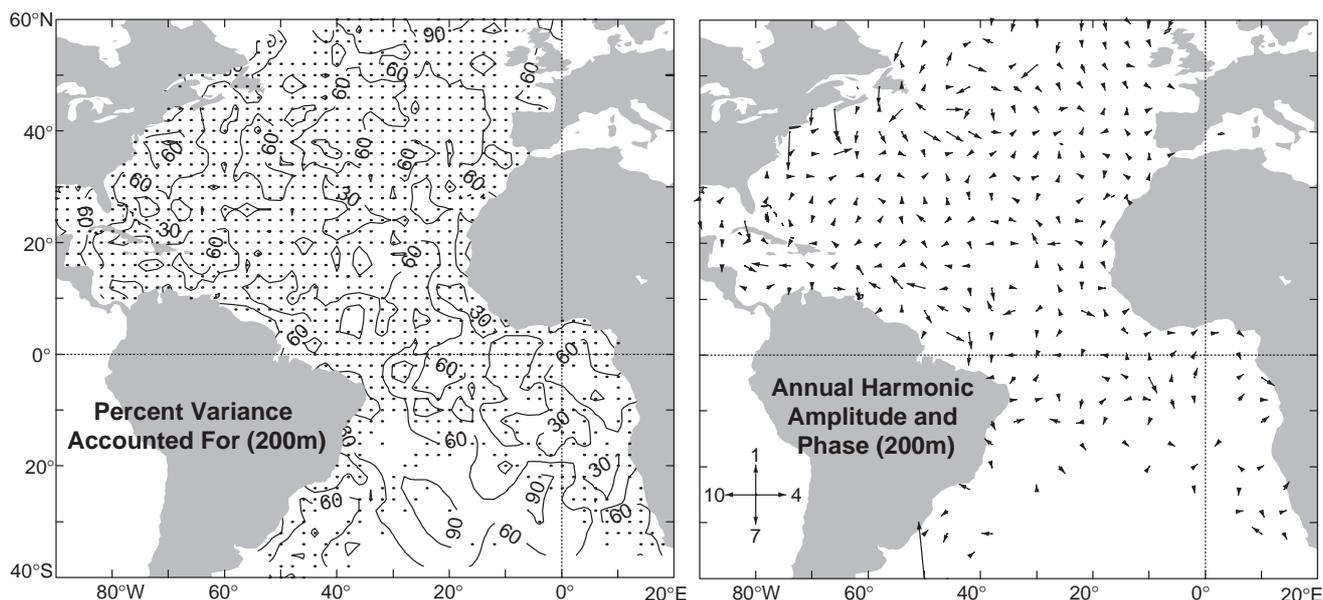


Figure 3. Left panel: Percent variance accounted for by the annual harmonic of the bi-monthly time-series available on the 2° of latitude by 2° of longitude grid for 200 m. The entire field is contoured by a machine software package, thus contours appear in data-void areas. Right panel: The amplitude and phase of the annual harmonic are given as arrows with lengths proportional to the amplitude of this harmonic and direction proportional to phase. Inset arrows provide the scale and have a length of 2°C and phase in months indicated as 1 = 1 January; 4 = 1 April, etc. Arrows are only given for those grid points with more than three bi-monthly values of data.

to 90% at the sea-surface (not shown). At this depth, large amplitudes are observed in the north-western Atlantic and in the western tropical Atlantic.

The representativeness of the WOCE years relative to the long-term XBT climatology is given in Fig. 4 (page 21). Anomalies at each 2° by 2° grid point with data are computed and presented for 3 years as data points that are superimposed on the mean temperature distributions. Those anomalies greater than one standard deviation different than the long term mean are highlighted. At 200 m, in general, the western subtropical gyre of the North Atlantic during the WOCE years experienced anomalously high temperatures. There is some suggestion of negative anomalies to the north of these positive anomalies.

Summary

Although sampling is uneven in both time and space, XBT data provide a valuable resource to study upper ocean temperature variability. For example, in the Atlantic, the XBT data have been combined with other temperature data to study long time changes in the North Atlantic (e.g., Levitus et al., 1994). The XBT data have also been used independently for similar purposes. In the tropics, Houghton (1991), for example, used these UOT data to study interannual variability. Similar studies using XBT data have also been conducted in the Pacific and Indian Oceans. Figures presented in this report will provide a basis for interested investigators to determine if sufficient data exist to assist in their research activities.

Acknowledgements

The efforts of Ms Yeun-Ho Daneshzadeh and Mr Mike Minton in establishing and running the Atlantic UOT DAC are gratefully acknowledged. Support is provided by NOAA's Office of Global Programs.

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The XXII General Assembly of the International Union of Geodesy and Geophysics

The twenty-second General Assembly of IUGG (International Union of Geodesy and Geophysics) will be held **18–30 July 1999 at the University of Birmingham, UK**. In order to maintain the high profile of WOCE you are strongly encouraged to submit abstracts.

Sessions of interest to WOCE scientists are:

- P07: Stable isotopes and trace substances: their use in oceanography and climate research on various timescales.
Convened by Peter Schlosser.
- P12: Global Water Mass Analysis.
Convened by Matt Tomczak.
- P13: Dynamics of sea ice and ocean polar seas.
Convened by Peter Wadhams.
- JSP05: Data assimilation in meteorology and oceanography.
Convened by Paola Rizzoli and Phillippe Courtier.
- JSP25: Ocean/Atmosphere variability and predictability.
Convened by Mike Davey, David Stephenson and Shoshiro Minobe.

- JSP49: Small-scale flow, turbulence and mixing.
Convened by Peter Baines.
- JSM18: Atmospheric and oceanic connections between the polar regions and lower latitudes including a session on boundary layers and surface wave and interfacial processes in honour of Henry Charnock.
Convened by Chuck Stearns and John Turner.
- JSM41: The contribution of satellite observations to global climate, ocean and terrestrial monitoring.
Convened by George Ohring.

Abstract deadline is 15 January 1999 and details of submission together with session outlines and the assembly programme can be found on the website

<http://www.bham.ac.uk/IUGG99/>

Copies of the third and preceding circulars are available from: IUGG99 Secretariat, School of Earth Sciences, University of Birmingham, Edgbaston, Birmingham, B15 2TT, UK, Tel: +44 121 414 6165, Fax: +44 121 414 4942, e-mail: IUGG99@bham.ac.uk.

Simulation of Interannual SST Variability in North Indian Ocean



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The annual cycle in the surface forcing of the north Indian Ocean due to monsoon reversal induces large variabilities in the thermodynamic characteristics of the upper ocean. These upper layer changes in turn might be influencing the seasonal and interannual variabilities in the Asian and Australian monsoon rainfall. The exact mechanism and processes involved though are not yet established. One of the major variables that are of great importance to such coupled variability in the air-sea interaction is the SST. Data scarcity has been a major factor for the restriction of such studies related to upper ocean variability and air-sea fluxes. As a result, most of the recent investigations rely heavily on model results. McCreary et al., 1993 (hereafter MKM) have used a 2.5 layer thermodynamic model of intermediate complexity to produce realistic climatology of SST and circulation in the upper Indian Ocean using climatological wind stress and heat fluxes. In the present study, the same model is used to simulate the interannual SST variability and to understand the relative importance of surface forcings on the evolution of SST anomalies.

The two active layers in the model interact with each other through entrainment and detrainment while conserving mass and heat of the total system. The first layer composed of two sub-layers, i.e. a well-mixed upper turbulent sub-layer and a non-turbulent fossil sub-layer. The upper sub-layer entrains or detrains water in a process in which the mixing is maintained by turbulence generated by both wind stirring and cooling at surface. The lower sub-layer is formed by the detrainment of water from the upper sub-layer and is kept isolated from the surface turbulence. However, it can be engulfed into the upper sub-layer either during the strong entrainment periods or upwelling regions where first layer becomes significantly shallow. There is also provision for detrainment of water from the first layer to the model second layer to conserve mass. There is no vertical shear in velocity field in the sub-layers of the first layer. The upper sub-layer temperature represents as the model SST.

The surface fluxes used in the model are derived from the monthly fields of the NCEP (National Center for Environment Prediction) reanalysis data (Kalnay et al., 1996). Net solar radiation (incoming fluxes

-outgoing fluxes), air temperature (at 2 m), specific humidity (at 2 m) and wind components (at 10 m) for the period 1982–1991 are used for flux computations. The model simulated SST fields are used for the heat flux computations instead of observed SSTs. Fields are obtained at model grid points by linear interpolation. The wind stress fields are computed from the wind fields with drag coefficient of 1.5×10^3 and air density of 1.2 kg/m^3 .

The model is first spun up for 10 years with the mean forcing that is obtained by 10 year averaging (1982–1991) of the surface flux fields. The numerical solutions reach a quasi-equilibrium state by the 10th year and these solutions serve as initial conditions for the interannual run in which the model equations are further integrated for 10 years with the interannually varying monthly surface forcings. The values of the variables at day 15 of each month represent the

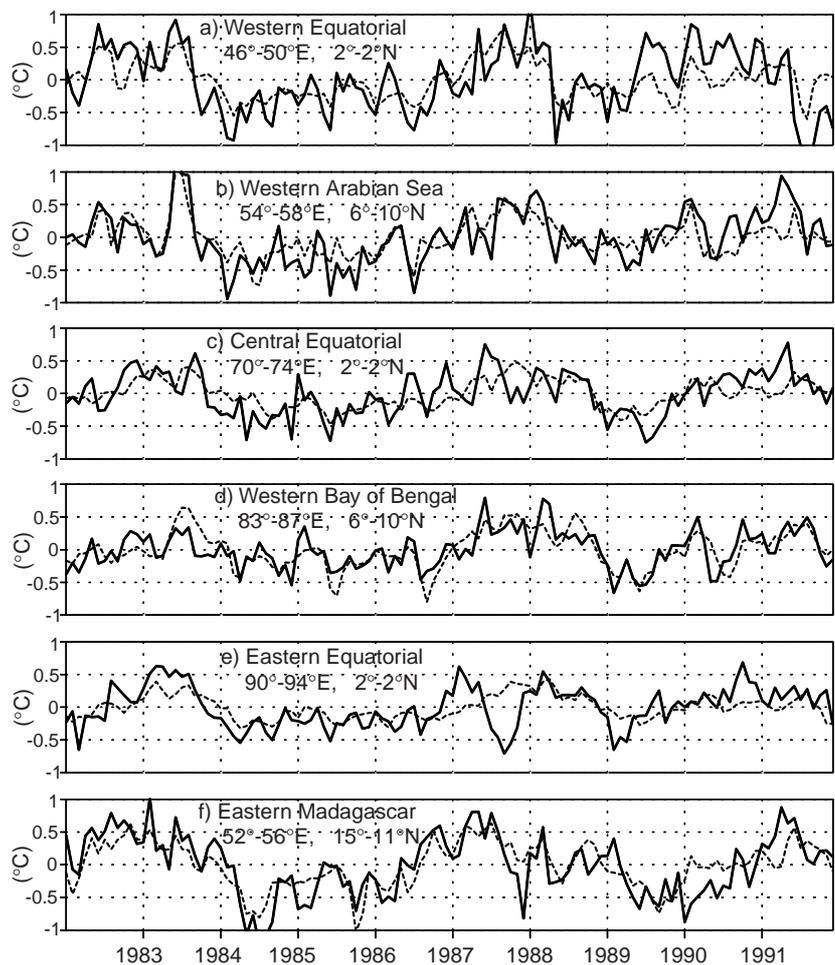


Figure 1. Time series of model SST anomalies (dashed lines) for six different regions. Superimposed are the observed SST anomalies (solid lines).

monthly values. Monthly anomalies are departures from the 10-year monthly average (monthly mean) of these variables. In general, these monthly mean model current and SST fields are found to be in good agreement with observed climatology.

The model simulates very well the interannual SST variability in most part of the model domain with slight exception of the central equatorial region and the Somali coastal region as understood from the RMS error and correlation analysis. In order to have a closer look at the model performance, the 10-year time series of the model SST anomalies are shown in Fig. 1 along with the Reynolds reanalysis SST (Reynolds and Marsico, 1993) anomalies in six small representative regions (5×5 box). Lack of high frequency sub-seasonal variability in the model forcing fields seems to affect the simulation of such high frequency variability in model anomalies. The time series of both model and data show three warm episodes during 1982–83, 1987 and 1990–91 and two cold episodes during 1984–86 and 1988–89. Except for the region in the west Bay of Bengal (Fig. 1d), in all other regions the model has poor

performance in the last two years of the model simulation; it is most likely because of the deficiency in the forcing fields. The model results, however, do not improve remarkably in a sensitivity experiment with FSU wind forcing and needs further investigation. The spatial patterns of the observed SST anomalies (not shown) in the month of July for three representative years show warm anomalies over most part of the basin in 1983 and 1987 and cold anomalies in 1985. A dipole structure in SST anomalies is also evident with opposite signs in the south-eastern part of the basin. Except for some detailed features found in the observed anomalies, the general spatial patterns of cold and warm anomalies are well simulated by the model.

In general, the SST anomalies are produced by the anomalies in surface forcings and anomalous horizontal and vertical advectations. Two forcings, the wind stress and the thermal convection at surface mainly drive the model ocean. In general, the SST anomalies are found to be in phase (opposite phase) with the corresponding anomalies of total heat flux (wind stirring). In a sensitivity experiment when the interannual heat flux is replaced with the

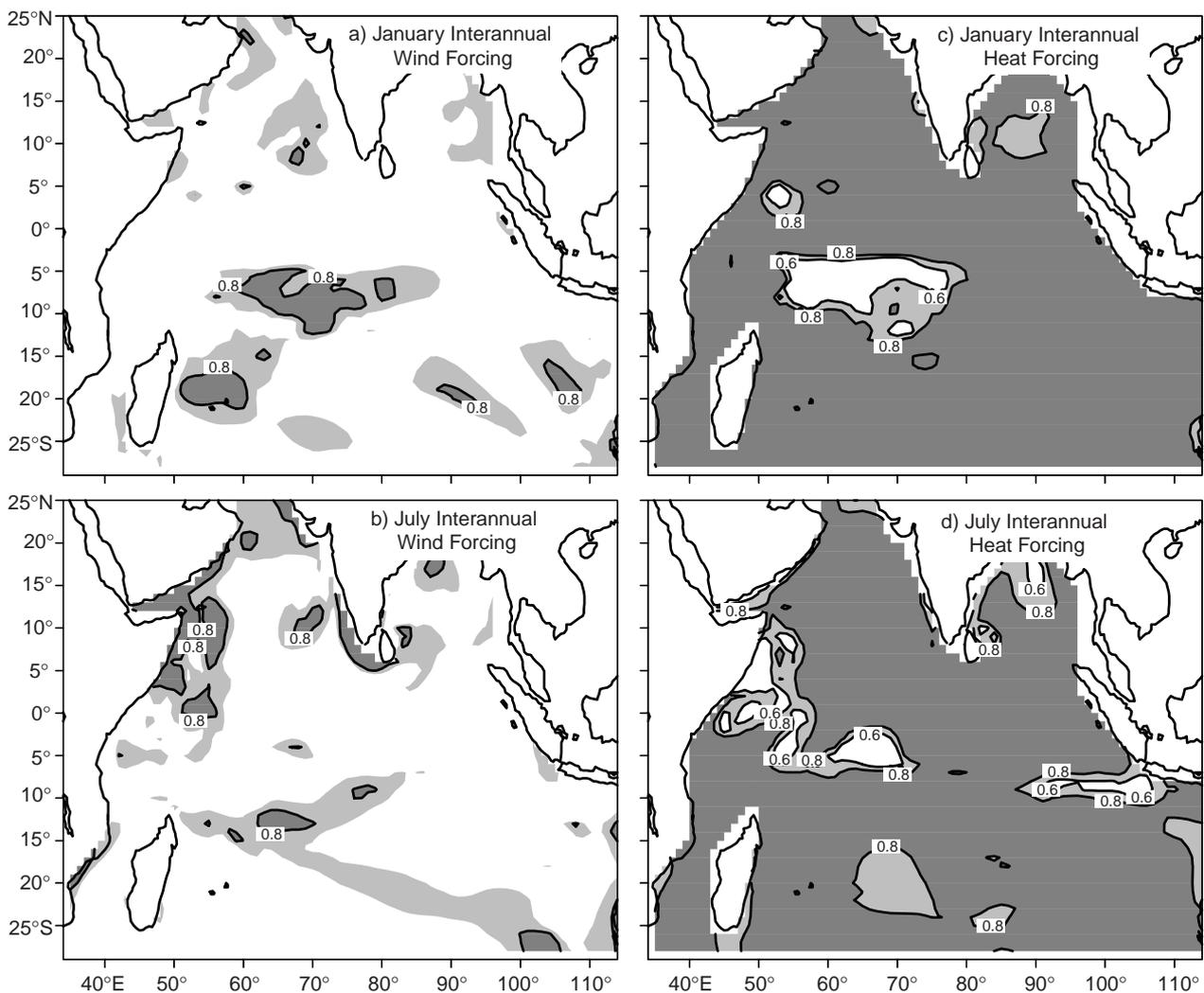


Figure 2. SST anomaly correlation between the control run and the interannual wind experiment for January (a) and July (b) and correlation between control run and interannual heat experiment for January (c) and July (d). Values >0.6 are shaded.

climatological (10-year averaged) heat flux, the phase is reasonably simulated. However, the amplitudes of the SST anomalies are found 50–60% less than the control run. In the second experiment the interannual wind forcing was replaced by the climatological (10-year averaged) wind forcing, while retaining the interannual heat forcing. The phase and amplitude of the SST anomalies in this case are remarkably similar to the control run indicating the dominant role of the heat forcing in the evolution of the SST anomalies.

The spatial patterns of the influence of the wind and heat forcings are illustrated in the process index (Fig. 2) obtained by taking the composites of correlation separately between the two sensitivity experiments and the control run for January and July. Comparison of the two panels again suggests the dominating influence of heat forcing over large regions (note correlation >0.8 in the heat forcing case). Also we note, the general correspondence between the regions of less correlation in heat forcing case and the high correlation in the wind forcing case. This is particularly the case in the Somali coastal region.

Radiative flux and the latent heat flux are the two major components of heat flux anomaly, sensible heat flux being a minor contributor. The role of the two major heat flux components varies from region to region and with time. Further analysis is being carried out to understand their role both in space and time. Two other sensitivity experiments suggest minor roles of the anomalous advections in the evolution of the anomalies, away from the coasts.

The time-longitude cross section of the observed (Reynolds) and model SST anomalies along the equator shows warm anomalies up to 0.5°C during 1982–83, 1987 and 1990–91. The warm anomaly bands seem to have a slight upward tilt to the east. Further, comparison of NINO-3 SST anomalies with the equatorial Indian Ocean SST anomalies for the period 1950–95 from the reconstructed historical SST data sets (Reynolds and Smith, 1994) suggest a general correspondence in the phase of the warm events with Indian Ocean having a lag of 3–4 months (Fig. 3). Therefore, the Indian Ocean also experience El-Niño like episodes (of a smaller amplitude) related to those in the Pacific. We note that the warm episodes in the model are not produced in the interannual wind forcing case and very

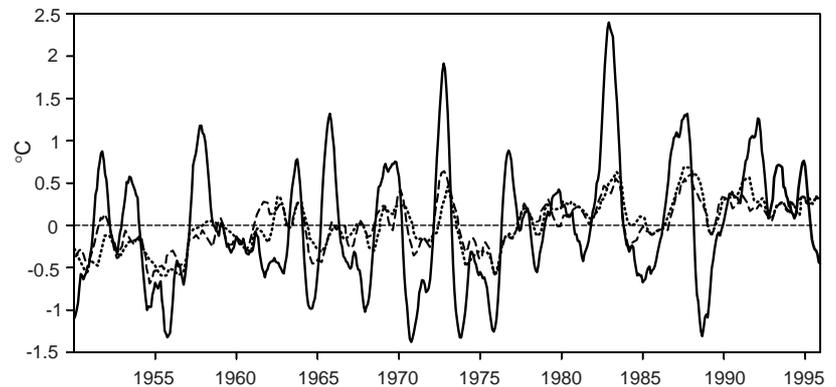


Figure 3. SST anomalies over NINO-3 region (solid lines), the east equatorial Indian Ocean (dotted line) and equatorial Indian Ocean (dashed line).

well reproduced in the interannual heat forcing case; again suggesting the importance of heat forcing in the evolution of these interannual events in the Indian Ocean. This shows quite a contrast to the Pacific phenomena. These results put forward necessity for high performance atmospheric models to produce realistic heat fluxes in coupled modelling studies of the region.

Acknowledgement

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Developments in Ocean Models Used for Climate Studies*

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The NCAR Climate System Model

The global ocean component of the NCAR Climate System Model is documented in Gent et al. (1998). Three main improvements were made compared to the previous generation of ocean climate models that used horizontal tracer diffusion and strong restoring to observations. The first was the replacement of horizontal tracer diffusion by the eddy parameterisation scheme of Gent and McWilliams (1990), that is much more clearly described in Gent et al. (1995). The second was the inclusion of the K-Profile Parameterisation scheme for the upper boundary layer of the ocean due to Large et al. (1994). The third was replacing strong restoring by bulk forcing, which is most thoroughly described in Large et al. (1997). Figure 2 of this paper summarises the improvements made by these three changes in the volume-averaged potential temperature and salinity in equilibrium solutions of a coarse ($3.6^\circ \times 3^\circ \times 25$ levels) non-eddy-permitting model.

Most of the improvement in the potential temperature distribution is due to the new eddy parameterisation, and most of the salinity improvement is due to the bulk forcing. The improvement due to the eddy parameterisation is, in part, due to elimination of false diapycnal mixing by horizontal diffusion. The strong restoring boundary condition on temperature is a reasonable approximation to the bulk forcing, because the non-solar terms do have the character of restoring terms. However, the correct fresh water boundary condition does not act like a restoring term at all, and a strong restoring boundary condition gives very unrealistic patterns compared to observations of evaporation minus precipitation. However, it is important to note that our bulk forcing for salinity does have a restoring to observations term, but with a restoring time on the order of a year. If this weak restoring term is set to zero, then the ocean simulation slowly, but surely, drifts away from a Levitus initial condition, and so would give a very unrealistic equilibrium solution. Thus realistic, ocean alone equilibrium solutions still need a weak restoring term in the fresh water boundary condition. However, I believe that global, non-eddy-permitting ocean solutions with horizontal tracer mixing and strong salinity restoring should be a thing of the past.

In a 300 year fully coupled run, the NCAR Climate System Model has a stable surface temperature distribution with no heat flux correction, see Boville and Gent (1998). However, there is no fresh water flux correction either, and

this leads to a considerable drift in the ocean salinity distribution over the 300 years, see Bryan (1998). I believe this salinity drift rate will be much reduced in the next version of the Climate System Model that will have improvements to all components. However, I anticipate that this improved coupled model will not have an equilibrium solution consistent with today's climate without a small fresh water flux correction. The reason is that precipitation is one of the most difficult fields for the atmospheric component to predict, and is very sensitive to changes in parameterisations and resolution. In addition, as noted above, small discrepancies between the fresh water forcing term applied, and what the ocean model requires, lead to large errors in the equilibrium ocean solution. I believe that eliminating the need for small fresh water flux corrections in coupled climate models is a much more challenging objective than eliminating the need for heat flux corrections.

Climate modelling at other centres

Support for the conclusions drawn in the section above has recently emerged from two other centres that concentrate on coupled climate modelling. The first is the Hadley Centre that now also has a fully coupled simulation with no flux correction that has little drift in the surface temperature distribution, (Roberts and Wood, pers. comm.). Their ocean component has a resolution of $1.25^\circ \times 1.25^\circ \times 20$ levels, uses the Visbeck et al. (1997) version of the eddy parameterisation and a vertical mixing scheme that is Kraus-Turner plus a simplified KPP scheme. Second, the CSIRO group at Aspendale have used the eddy parameterisation scheme in their ocean component, which has a resolution of $5.6^\circ \times 3.2^\circ \times 21$ levels. They find that the heat flux corrections required for a stable coupled integration with this version are much smaller than those from a model with the same resolution that used horizontal tracer diffusion. The reduced heat flux corrections are especially dramatic in the southern hemisphere near the Antarctic Circumpolar Current, see Hirst et al. (1998).

Future developments in ocean models

There are several developments that are being worked on by various climate modelling groups. There seems to be a general consensus that they are important, and will, hopefully, contribute towards better ocean component solutions. A partial list of these developments is:

- (a) Enhanced meridional resolution and smaller horizontal viscosity at the equator. This produces much more realistic equatorial currents at a modest increase in

* This is an extended abstract of a talk given at the WOCE/CLIVAR Workshop on Ocean Modeling for Climate Studies, which was held at NCAR on 10–13 August 1998. The talk was divided into four sections.

computational expense.

- (b) Allowing the background mixing coefficients to be functions of space. This might be most important for the isopycnal mixing coefficient, because this parameterisation is supposed to represent mixing by mesoscale eddies, which is highly spatially variable.
- (c) Improving the upper boundary layer mixing parameterisation. How well does it represent all the ocean processes that it is supposed to represent; double diffusion for example?
- (d) Include an explicit bottom boundary layer code, especially in z-coordinate models. There are four schemes I know of: Beckmann and Doscher (1997), Gnanadesikan (1998), Killworth and Edwards (1998), and Song and Chao (1998). These schemes may well work better using the partial bottom cell scheme discussed in Adcroft et al. (1997).
- (e) Adding an explicit river run-off scheme. This allows for a better representation of the fresh water input, or forcing, of global ocean models. It is also a necessary step to a consistent balancing of the fresh water budget in a coupled climate model.

Resolution of future ocean models

In this section, I address the question of what should the resolution be of ocean models used for climate studies in 3 to 5 years time?

The first point to make is that a recent study at Los Alamos and NCAR has shown that to “resolve” the ocean mesoscale field requires a horizontal resolution of 10 km or smaller. This study is summarised in an article by Bryan and Smith in this Newsletter. It consists of a series of North Atlantic simulations using progressively finer resolution, that were run out for a decade or two. Thus, the possibility of running “eddy-resolving” resolution for climate models is out of the question for the foreseeable future.

Thus, the question to be addressed is should non-eddy-permitting (1° or coarser) or eddy permitting resolution (finer than 1°) be used? I think our current eddy parameterisations have been much more thoroughly tested in the non-eddy-permitting regime. However, there is no doubt that finer horizontal resolution allows faster western boundary currents, for example, and can resolve narrow straits and topographic features much better. The difficult question is then “How important are these more realistic features in improving the ability of the ocean model to realistically capture climate variability compared to the rather large increase in computational cost”? Also, in terms of computational cost, how does one balance the relative merits of fewer coupled runs with higher resolution components against lower resolution components that allow more sensitivity experiments and ensembles of climate runs to be performed?

These are difficult questions to answer, but my opinion is that we will learn more about the climate system by running more coupled simulations rather than running fewer with eddy-permitting ocean resolution. I believe it may be more important to resolve narrow straits better and have faster western boundary currents than to have global eddy-permitting resolution. This could be achieved by using irregular, or adaptive, horizontal grids that allow finer resolution in predetermined locations. This would result in a much more modest increase in computational cost than using a globally uniform finer grid. I think that the standard ocean model resolution used for climate work in 3 to 5 years will be 1° , with finer resolution at some specific locations. I believe that this is the resolution that WOCE/CLIVAR development work should be aimed at.

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Modelling the North Atlantic Circulation: From Eddy-Permitting to Eddy-Resolving



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The first goal of WOCE is to develop models useful for predicting climate change and to collect the data necessary to test them. The WOCE Scientific Plan (WCRP, 1986) describes the iterative process of model development, tuning, and testing against observations. In particular, it identifies Goal 1 as being:

“... concerned with one major iteration of this cycle: the rigorous testing of eddy resolving, thermodynamically active models on the scale of an ocean basin, including at least a cyclonic and anti-cyclonic gyre...”

Implicit in this statement is the assertion that eddy-resolving models would be necessary to meet the WOCE requirement of being useful for predicting climate change, though the definition of “eddy resolving” is left unspecified. During the WOCE era, the computational resources available to the ocean modelling community have increased from a level that allowed basin scale, thermodynamically active, simulations with horizontal resolutions of 35–40 km and 20–30 vertical levels to the present capability of simulations at 5–10 km horizontal resolution and 40–50 vertical levels. Simulations across this whole range of resolutions have been referred to as “eddy-resolving”, though their success in representing the observed distribution and amplitude of mesoscale variability has frequently not been terribly good (Fu and Smith, 1996; McClean et al., 1997; Maltrud et al., 1998). Further, the simulation of ocean heat transport has in some of these cases been poorer than in low resolution (100 km or coarser) models that incorporate modern parameterisations of mesoscale eddies (Böning et al., 1995). Evaluating the role of resolution in the successes and failures of these simulations has been difficult due to the wide variety of other modelling factors that have been varied simultaneously.

The goal of this study is to systematically evaluate the resolution dependence of the simulation of both mesoscale variability and climatically relevant integral quantities such as heat transport in a basin-scale ocean model. Three experiments have been carried out with the POP (Dukowicz and Smith, 1994) primitive equation ocean model that differ only in their horizontal resolution and dissipation parameters. A Mercator grid is used with equatorial resolution of 0.1°, 0.2° or 0.4°. The domain includes the Atlantic Ocean from 20°S to 73°N latitude, the Gulf of Mexico, Caribbean Sea and western Mediterranean Sea. Only the 0.1° case has grid spacing smaller than the first baroclinic mode Rossby radius of deformation over the entire domain. All cases have 40 vertical levels with near surface resolution of 10 m. Each experiment is forced with daily wind stress based on ECMWF surface analyses for the period 1985 to 1996, the Barnier et al. (1995)

climatological monthly heat flux (including a penetrative solar flux), and restoring of surface salinity to the Levitus climatology. A biharmonic formulation of subgrid-scale viscous and diffusive dissipation, acting along geopotential surfaces, is used in all cases. The eddy viscosity and diffusivity vary spatially with the cube of the grid size, both on each grid and between grids. Each experiment is initialised at rest with the Levitus climatological temperature and salinity and integrated for 15 years.

In this short note we present two comparisons of the results of these experiments against observations. The first is a comparison of variability of sea surface height against altimetric observations. The second is a comparison of poleward ocean heat transport against various direct and indirect measures. More extensive results are available in a forthcoming manuscript (Smith et al., 1999) and on the World Wide Web at:

<http://www.cgd.ucar.edu/oce/bryan/woce-poster.html>

The r.m.s. sea surface height variability for each of the three model experiments and that observed by the TOPEX/POSEIDON and ERS-1/2 altimeters (Le Traon and Ogor, 1998) is shown in Figs. 1a–d (page 22). The distribution and amplitude of variability change quite dramatically, both qualitatively and quantitatively, across this range of resolution. The 0.2° and 0.4° cases have similar (and similarly unrealistic) geographical distributions of variability and differ primarily in amplitude. At 0.1° resolution the model simulation of both the distribution and amplitude is in good agreement with the altimetric observations over most of the domain. Notable points of improvement over the lower resolution experiments include the Gulf Stream between Cape Hatteras and the Grand Banks, the North Atlantic Current in the “Northwest Corner” east of Newfoundland and the Azores Current. In each of these examples the improvement in the simulation of variability accompanies an improvement in the simulation of the mean flow. There appears to be a significant regime transition in the realism of the simulations of both the time mean flow and variability between 0.1° and 0.2° resolution. In the highest resolution case the Gulf Stream separates from the coast at Cape Hatteras as a jet, while in the 0.2° and 0.4° cases the separation point is displaced to the north and a strong, tight anticyclone develops offshore. Surface velocities in the jet core of the highest resolution case reach 2 m/sec and the stream-coordinate average downstream transport reaches 150 Sv, in quantitative agreement with observations (Johns et al., 1995). The meander envelope and the maximum in sea level variability occur in the vicinity of the New England Seamount Chain. In the 0.1° case the North Atlantic Current (NAC) turns north and slightly west as it passes the Grand Banks of Newfoundland

to form the “Northwest Corner”, while in the two lower resolution cases the NAC follows a more eastward course towards the mid-Atlantic Ridge. In consequence, in the 0.1° case the variability maximum in the subpolar region occurs in the Labrador basin as observed, rather than in the Irminger basin as in the lower resolution cases. In addition, a distinct frontal zone with time-mean eastward surface flow of 0–12 cm/sec, a net eastward transport of 9.6 Sv, and a local maximum in sea surface variability emerges along the Azores Front at 35°N in the eastern basin at the highest resolution, but is absent in the lower resolution cases. In addition to the favourable comparisons of the geographical distribution and amplitude of SSH variability, the 0.1° case shows good quantitative agreement with other measures of mesoscale variability such as: wavenumber-frequency spectrum of SSH anomalies compared to altimetric estimates in the Gulf Stream region, near-surface EKE compared to drifter based statistics, and vertical profiles of EKE compared to current meter observations.

The net meridional heat transport for each of the three experiments is shown in Fig. 2. An estimate based on an atmospheric residual calculation (Trenberth, 1998) and the inverse model estimate of MacDonald and Wunsch (1996) are shown for comparison. For latitudes south of 35°N , only the 0.1° experiment heat transport falls within the uncertainty range of the Trenberth (1998) estimate, and all three model results fall below the inverse model calculation at 11°S and 11°N . All three model results exceed the Trenberth (1998) observational estimate poleward of 45°N , though they all fall within the uncertainty of the northernmost estimate of MacDonald and Wunsch (1996) at 48°N . Across this range of resolutions the meridional heat transport increases monotonically with increasing resolution with no indication of convergence of the solution; at most latitudes the difference between the 0.1° and 0.2° cases is as large or larger than that between the 0.2° and 0.4° cases. Despite the fact that there is a significant difference in the level of mesoscale variability between the three experiments (Fig. 1), the majority of the increase in heat transport with resolution is due to changes in the heat transport by the time mean flow, not the eddy heat transport. Between the equator and 32°N the eddy heat transport is southward in all three cases, but the transport by the time mean flow and the total transport become increasingly northward.

The results of these experiments have several implications for the prospect of meeting WOCE Goal 1. At the highest resolution, the solutions obtained are in quantitative agreement with many observed measures of the North Atlantic general circulation and its variability. This result is consistent with our current understanding of the relationship between the dominant

scales of mesoscale variability in the ocean and the Rossby radius of deformation (Stammer, 1997) and the fact that only the highest resolution case resolves this scale over the full domain. Thus, a resolution of around 0.1° appears to be required for a simulation to be “eddy-resolving”. With this experiment and those planned for the future we can begin in earnest the iterative process of model testing described in the WOCE plan. Further, this experiment and others at similar resolution will provide a rich and valuable data set with which to investigate the dynamical balances of the general circulation, test sub-grid scale parameterisations for lower resolution models, assist in designing future observational strategies, etc.

On the other hand, our lower resolution cases are unable to achieve ocean heat transports within the uncertainty of the observational estimates and we cannot demonstrate convergence of the numerical solutions within a range of resolutions that will become practical for coupled climate modelling for the foreseeable future. However, such high resolution may not be necessary to make progress toward WOCE Goal 1. Experience with the NCAR Climate System Model (Boville and Gent, 1998) and other recent coupled models show that it is possible to realistically simulate poleward ocean heat transport at low resolution with adequate parameterisation of the effects of mesoscale eddies. The results of the 0.2° and 0.4° cases described here are an indication that the effects of mesoscale eddies on the time mean flow will still need to be parameterised in this “eddy-permitting” resolution regime. As stated above, the availability of truly “eddy-resolving” simulations will help to facilitate the development of suitable parameterisations.

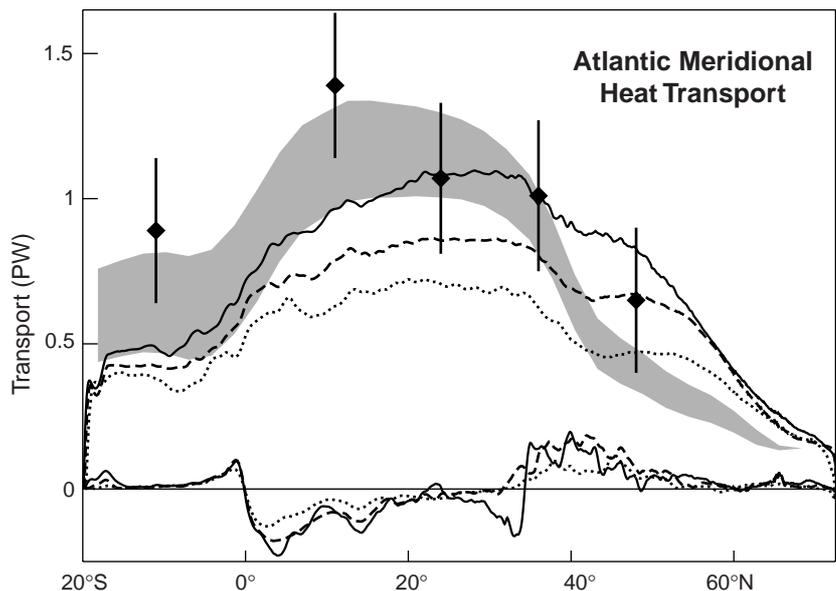


Figure 2. Net meridional heat transport (upper curves) and transport by the time varying flow (lower curves) for the North Atlantic models at 0.1° (solid), 0.2° (dashed), 0.4° (dash-dotted) resolution. The estimate of Trenberth (1998) is indicated by the shaded region and the inverse model results of MacDonald and Wunsch (1996) are indicated by the diamond symbols and vertical bars.

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Second Summer School on Inverse Methods and Data Assimilation

19–30 July 1999, College of Oceanic and Atmospheric Sciences, Oregon State University, Corvallis, OR, USA

The College of Oceanic and Atmospheric Sciences at Oregon State University will conduct a Summer School on Inverse Methods and Data Assimilation in July 1999. The programme will closely follow the 1997 Summer School, with a two-week programme of three daily lectures, plus introductory computing exercises. In 1999 there will in addition be a Workshop on Advanced Implementation. It will be held during the second week (28–30 July), and may be attended without participating in the rest of Summer School. The Workshop is directed at returning participants and experienced practitioners, but first-time participants who have completed the computing exercises may also attend, as space permits.

The lecture syllabus is given at the end of this announcement. Class notes will be distributed at the commencement of the Summer School. The theoretical material presented in lectures will be supported by detailed discussions of major applications in oceanography and meteorology involving real data.

The computing laboratory will house about 30 fully-networked UNIX workstations reserved for the Summer School. Ambitious students may wish to use the CM-500e Connection Machines at the College.

A block of single-occupancy rooms with own or pair-shared bathrooms has been reserved at the College Inn, an upperclassman hall of residence adjoining the OSU campus. Funds for travel, accommodation and meals are available for US-based participants. Travel support for foreign-based participants is being sought.

The 1999 Summer School is sponsored by ONR, NASA, NSF, NOAA Sea Grant and OSU.

For further information please contact the Summer School secretary, Mrs Florence Beyer (fbeyer@oce.orst.edu).

Syllabus

Forward Models, Inverse Models, Solving the Euler-Lagrange Equations Using Representers, Regularity of the Inverse Estimate, Some Limiting Choices of Weights: ‘Weak’ and ‘Strong’ Constraints, Geometrical Interpretation, Statistical Interpretation: The Relationship to ‘Optimal Interpolation’, The Reduced Penalty Functional, General Measurement, Accelerating the Representer Calculation, Array Modes, The Primitive and Quasi-geostrophic Equations, Generalized Inversion of a Quasigeostrophic Model, Smoothing Norms, Covariances and Convolutions, Gradients, The Sweep Algorithm, Sequential Estimation and the Kalman Filter, The Kalman Filter: Statistical Theory, Ill-Posed Problems, Nonlinear and Nonsmooth Estimation, Importance Sampling, Ensembles and Simulated Annealing, Posterior Errors.

Staff

Director:	A. Bennett
Summer School lecturers:	A. Bennett, J. Mead
Workshop chairmen:	B. Chua, H.-E. Ngodock

Eddy Parameterisations in the Stochastic Theory of Turbulent Fluid Transport



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Mesoscale eddies play an important role in determining the character of the mean circulation in the ocean. Only very recently has increased computing power made possible basin-scale eddy-resolving ocean general circulation models (OGCMs) with sufficient spatial resolution to quantitatively reproduce basic features of the mean circulation such as the separation of western boundary currents and the shape and geographical extent of the subtropical and subpolar gyres (see the article in this issue: “Modelling the North Atlantic Circulation: from Eddy-Permitting to Eddy-Resolving”). Unfortunately, these models are so expensive to integrate that it is only possible at present to conduct 10–20 year simulations. Eddy-resolving climate simulations with full 3-dimensional OGCMs will not be computationally feasible for the foreseeable future. However, recent subgridscale parameterisations of mesoscale processes have shown promise in improving the ability of non-eddy-resolving models to simulate the meridional heat transport and other quantities relevant to climate variability. The Gent-McWilliams parameterisation (Gent and McWilliams, 1990, hereafter GM90) of eddy-induced tracer transport is based on downgradient diffusion of thickness along isopycnal layers to mimic the effects of mesoscale eddies generated through baroclinic instability. In addition to improving model heat transport, the parameterisation has other beneficial effects such as reducing spurious diapycnal mixing in level models. This parameterisation has a theoretical basis in the stochastic theory of turbulent transport in adiabatic stratified flow, recently developed by Dukowicz and Smith (1997, hereafter DS97). This article discusses recent progress in the development of stochastic closure schemes for ocean mesoscale turbulence.

The stochastic theory

Stochastic or random-walk theories underlie semi-empirical models of turbulent diffusion involving downgradient or Fickian diffusion. They are the basis for simple one-point and mixing-length theories and form the foundation for more elaborate closures. The stochastic theory is attractive because it makes minimal assumptions about the character of the turbulence: it assumes only that the turbulent flow is random (meaning the fluid-parcel displacements can be described by a probability distribution function or pdf) and Markovian (meaning fluid parcels forget all but their most recent past history). These assumptions will apply to fluid turbulence if it is not strongly inhomogeneous and the parcel trajectories are not correlated over very long times. Until recently it was not possible to apply the stochastic theory to ocean mesoscale turbulence because the theory assumed incompressible flow or a divergence-free velocity field. In the interior adiabatic ocean mixing is believed to

be confined to isopycnal layers where the horizontal velocity field is divergent, and hence a compressible-type theory is needed, where isopycnal layer thickness plays the role of density. In the standard incompressible theory (Monin and Yaglom, 1971, hereafter MY) a semi-empirical diffusion equation (SEDE) for the transport of a passive tracer is derived from the Fokker-Planck equation (the evolution equation for the pdf) using a suitable average of the pdf over an initial tracer distribution. The incompressible SEDE consists of advection of the mean tracer distribution with the Eulerian-mean velocity and downgradient diffusion with a turbulent diffusion tensor \mathbf{K} . In DS97 it was shown that by making the *same* assumptions as in the standard incompressible theory, an SEDE can be derived for compressible-type flow which also involves downgradient diffusion, but in addition it predicts tracer advection with the Eulerian-mean velocity *plus* an eddy-induced transport velocity involving downgradient diffusion of the mean density (or layer thickness) with the same diffusivity tensor \mathbf{K} . The layer continuity and tracer transport equations in this theory are given in isopycnal coordinates by:

$$\partial_t \hat{\phi} + (\mathbf{U} + \mathbf{U}^*) \cdot \nabla \hat{\phi} = \frac{1}{h} \nabla \cdot \bar{h} \mathbf{K} \cdot \nabla \hat{\phi} \equiv \mathfrak{R}(\hat{\phi}) \quad (1)$$

$$\partial_t \bar{h} + \nabla \cdot \bar{h} (\mathbf{U} + \mathbf{U}^*) = 0 \quad (2)$$

$$\mathbf{U} = \mathbf{v} - \nabla \cdot \mathbf{K} \rightarrow \bar{\mathbf{u}} \quad (3)$$

$$\mathbf{U}^* = -\frac{1}{h} \mathbf{K} \cdot \nabla \bar{h} \quad (4)$$

where ϕ is a passive tracer and h is the layer thickness. The bar denotes an ensemble-mean taken at fixed density in isopycnal coordinates, and the caret denotes a thickness weighted ensemble-mean: $\hat{\phi} = \bar{h}\phi / \bar{h}$. Thickness-weighted averages are needed in the compressible-type theory to insure conservation of mass and tracers between isopycnal layers. The r.h.s. of (1) is a diffusion operator that mixes ϕ along isopycnal surfaces. In the limit of constant thickness (incompressible-type flow) Eq. (2) becomes $\nabla \cdot \mathbf{U} = 0$ and Eq. (1) reduces to the standard SEDE. The unknowns in these equations are the Lagrangian-mean velocity \mathbf{v} and the diffusivity tensor \mathbf{K} , which are given, respectively, by the first and second moments of the pdf with respect to particle displacements. \mathbf{K} is a symmetric positive-definite 2×2 along-isopycnal mixing tensor. In the incompressible case the standard hypothesis is that the transport velocity $\mathbf{U} = \mathbf{v} - \nabla \cdot \mathbf{K}$ is given by the Eulerian-mean velocity $\bar{\mathbf{u}}$ (MY; we refer to this as the “MY hypothesis”). In DS97 it is shown that this is also the simplest hypothesis in the compressible case. The term $\nabla \cdot \mathbf{K}$ is associated with the tendency of a cloud of particles in inhomogeneous turbulence

to move or be dispersed even in the absence of mean flow. The eddy-induced transport velocity \mathbf{U}^* is proportional to a downgradient thickness flux, as in the GM90 parameterisation. In the limit of an isotropic scalar diffusivity that is depth-independent, these equations are equivalent to the GM90 parameterisation with a single diffusivity for both thickness and tracers.

How are the quantities in Eqs. (1–4) related to the eddy statistics? Performing a Reynolds' decomposition of the variables into mean and fluctuating components: $\phi = \bar{\phi} + \phi' = \hat{\phi} + \phi''$, the ensemble-mean continuity and transport equations are

$$\partial_t \hat{\phi} + \left(\bar{\mathbf{u}} + \frac{1}{h} \overline{\mathbf{u}'h'} \right) \cdot \nabla \hat{\phi} = -\frac{1}{h} \nabla \cdot \bar{h} < \mathbf{u}'' \phi'' > \quad (5)$$

$$\partial_t \bar{h} + \nabla \cdot (\bar{\mathbf{u}}\bar{h} + \overline{\mathbf{u}'h'}) = 0 \quad (6)$$

These involve two correlations: the eddy thickness flux $\overline{\mathbf{u}'h'}$ and the eddy tracer flux $< \mathbf{u}'' \phi'' >$, where brackets $< \dots >$ also denote a thickness-weighted average. Comparing Eqs. (1–4) with (5–6) it can be shown that these are given by

$$\overline{\mathbf{u}'h'} = -\mathbf{K} \cdot \nabla \bar{h} + \mathbf{k} \times \nabla \psi_h \quad (7)$$

$$\bar{h} < \mathbf{u}'' \phi'' > = -\bar{h} \mathbf{K} \cdot \nabla \hat{\phi} + \mathbf{k} \times \nabla \psi_\phi - \hat{\phi} \mathbf{k} \times \nabla \psi_h \quad (8)$$

where ψ_h and ψ_ϕ are scalar gauge fields determined by the rotational (divergence-free) components of the thickness and tracer fluxes. The stochastic theory makes no predictions about the form of the rotational fluxes, so these gauge fields are left unspecified. However, rotational fluxes are generally comparable or larger in magnitude than their corresponding divergent components. The presence of rotational fluxes makes it particularly difficult to extract quantities such as diffusivities from turbulent fluxes derived from observations or model output.

The momentum equation and conservation of potential vorticity

In the stochastic theory the transport equation (1) only applies to quantities ϕ that are conserved following fluid parcel trajectories. Since momentum is not conserved along trajectories, the theory cannot be directly applied to the momentum equation. However, if the potential vorticity is conserved in the interior, then Eq. (1) also applies with $\phi = q$, where $q = (f + \zeta)/h$ and ζ is the relative vorticity. This provides a constraint on part of the Reynolds correlation in the ensemble-mean momentum equation

$$\begin{aligned} \partial_t \bar{\mathbf{u}} + \bar{\mathbf{u}} \cdot \nabla \bar{\mathbf{u}} + f \mathbf{k} \times \bar{\mathbf{u}} + \nabla \bar{M} \\ = -\overline{\mathbf{u}' \cdot \nabla \mathbf{u}'} = -\nabla \overline{\mathbf{u}' \cdot \mathbf{u}' / 2} - \mathbf{k} \times \overline{\mathbf{u}' \zeta'} \end{aligned} \quad (9)$$

$$\overline{\mathbf{u}' \zeta'} = \hat{q} \overline{\mathbf{u}' h'} + \bar{h} < \mathbf{u}'' q'' > \quad (10)$$

where M is the Montgomery potential. Equation (10) is an exact identity which relates the turbulent fluxes of relative vorticity, thickness and potential vorticity. Inserting Eqs.

(7–8) with $\phi = q$, the r.h.s. of (9) becomes

$$-\overline{\mathbf{u}' \cdot \nabla \mathbf{u}'} = -\nabla \chi + \mathbf{k} \times \mathbf{K} \cdot \nabla (f + \bar{\zeta}) \quad (11)$$

where $\hat{q} = (f + \bar{\zeta})/\bar{h}$, $\chi = \overline{\mathbf{u}' \cdot \mathbf{u}' / 2} - \psi_q$ and ψ_q is the gauge field associated with the rotational component of the relative vorticity flux: $\overline{\mathbf{u}' \zeta'} = -\mathbf{K} \cdot \nabla (f + \bar{\zeta}) + \mathbf{k} \times \nabla \psi_q$. Thus the turbulence closure problem for the ensemble-mean adiabatic primitive equations is reduced to the determination of one scalar field χ and the diffusivity tensor \mathbf{K} (the latter is also a single scalar field in the limit of horizontally isotropic turbulence). The requirement that the net change in mean kinetic plus potential energy is positive or zero imposes an integral constraint on χ , but it is otherwise unspecified in the stochastic theory. This closure is derived and discussed in more detail by Smith (1998, hereafter S98).

The planetary-geostrophic regime

A weak point in the derivation of this closure is the MY hypothesis. Dukowicz and Greatbatch (1998a, hereafter DG98) show that there is at least one regime of the flow, the planetary geostrophic (PG) limit, where this hypothesis must be modified. In that limit the potential vorticity is simply given by f/h and the eddy fluxes of thickness and potential vorticity are exactly related by Eq. (10) with $\overline{\mathbf{u}' \zeta'} \rightarrow 0$ on the l.h.s. DG98 use this relation to show that, without making the MY hypothesis the eddy-induced transport velocity is given in the PG regime by

$$\mathbf{U}^* = \frac{1}{\hat{q}} \mathbf{K} \cdot \nabla \hat{q} - \frac{1}{h\hat{q}} \mathbf{k} \times \nabla \psi_q \quad (12)$$

They then hypothesise that the gauge term $\nabla \psi_q$ vanishes, so the eddy-induced transport velocity is given by a diffusive flux of q rather than h , as in the linear-wave theory of Killworth (1997). In S98 it was shown that in the PG and quasi-geostrophic (QG) regimes the hypothesis that $\nabla \psi_q$ vanishes is equivalent to the MY hypothesis, and the time evolution of the tracer distribution will be identical for the two closure schemes. Note, however, that in (12) \mathbf{U}^* is singular near the equator where \hat{q} can vanish, while the approach based on Eqs. (1–4), (9), and (11) has the advantage that it is everywhere non-singular and thus can be used in global simulations.

Open questions

The stochastic theory has proven successful in constraining the form of the turbulence closure for adiabatic stratified turbulence and providing a theoretical foundation for the GM90 parameterisation, however, several important issues remain unresolved:

- (1) What is the nature of the boundary conditions for adiabatic flow near topography in the stochastic theory? Greatbatch and Li (1998) have investigated the effects of including the gradient of ocean depth H in downgradient parameterisations of the q -flux. A related question is how does this closure scheme match up with parameterisations of diabatic turbulent flow in the surface and bottom boundary layers?

(2) How valid is the MY hypothesis (or the equivalent gauge hypothesis of DG98), and how can it be tested in model simulations?

(3) What is the space and time dependence of \mathbf{K} and χ ? Can these be specified by a simple algebraic closure, or will it be necessary to develop a second order closure with prognostic equations for the time-evolution of these fields? From the formal definition of the diffusivity tensor in terms of the pdf (DS97, Eq. 7b) it is clear that \mathbf{K} should increase with the intensity of the turbulence, and vanish wherever the flow is not turbulent, so it is expected to have significant spatial variation. Several forms for the spatial dependence of a scalar diffusivity κ have been proposed in the literature. Recently Bryan et al. (1998) analysed the eddy-thickness flux in a global eddy-resolving model and found that the form of κ proposed by Visbeck et al. (1997) gave the best fit to the model statistics compared to other proposed forms. In their parameterisation $\kappa \propto R_1^2 / T$ where R_1 is the first baroclinic Rossby radius and T is the Eady time-scale for the growth of unstable baroclinic waves. The key feature of this form of κ that brings it into better agreement with the model statistics is that it is inversely proportional to the Coriolis parameter f . A QG scale analysis of the stochastic theory closure scheme (S98) suggests that κ scales like $R_o U L = U^2 / f$ (where R_o is the Rossby number and U, L are typical velocity, time scales) which has the same inverse dependence on f .

(4) What are the limits of applicability of the stochastic theory? Holloway (1997) has pointed out that the theory of statistical mechanics predicts certain types of rectified mean flows (e.g., the Neptune Effect) which cannot be associated with downgradient turbulent fluxes. Dukowicz and Greatbatch (1998b) have shown that mean-flow Fofonoff gyres which appear in the inviscid barotropic QG solutions are not consistent with a downgradient flux of potential vorticity. However, they found that if a viscous term of sufficient strength is present in the momentum equation then the Fofonoff gyres will be homogenised, implying that the q -flux is downgradient and the stochastic theory will apply. They predict the flow will have both inviscid and viscous regimes, whose manifestation depends on the relative magnitude of an inertial and a viscous timescale. A related problem is associated with the transfer of mean to eddy enstrophy. In the stochastic theory the transport equations for mean and eddy tracer variance have the simple form:

$$\hat{D}_t \frac{1}{2} \hat{\phi}^2 = -\nabla \hat{\phi} \cdot \mathbf{K} \cdot \nabla \hat{\phi} + \Re \left(\frac{1}{2} \hat{\phi}^2 \right) \quad (13)$$

$$\hat{D}_t \frac{1}{2} \langle \phi''^2 \rangle = +\nabla \hat{\phi} \cdot \mathbf{K} \cdot \nabla \hat{\phi} + \Re \left(\frac{1}{2} \langle \phi''^2 \rangle \right) \quad (14)$$

where $\hat{D}_t = \partial_t + (\bar{\mathbf{u}} + \mathbf{U}^*) \cdot \nabla$. The first terms on the r.h.s. are a local sign definite sink from mean to eddy tracer variance. Integrating these equations over the domain and assuming no fluxes of tracers through the boundaries, it follows that in a statistical steady state the volume integral of this conversion term must vanish. If turbulence is present in the final steady state then $\mathbf{K} \neq 0$, hence the conversion

term can only vanish if ϕ is constant in each layer. This makes sense for passive tracers: it means the turbulence homogenises the tracer distributions within each layer. In the case of $\phi = q$, the turbulence also acts to homogenise potential vorticity, but q will not in general obey the same no-flux boundary conditions as passive tracers, and the final state cannot have constant q everywhere in the layer. So there must be something else to balance the sink from mean to eddy enstrophy: it must either be exactly balanced by the boundary fluxes of q , which is unlikely, or there must be another term on the r.h.s. of (14 with $\phi = q$) which dissipates eddy enstrophy $\frac{1}{2} \langle q''^2 \rangle$. Such a term would naturally arise if there were a friction term in the momentum equation. The need for friction in the interior when the q -flux has a component down the mean gradient was first pointed out by Rhines and Holland (1979). This friction, which is presumably caused by internal breaking waves or other small-scale processes, evidently plays a crucial role the process of potential vorticity homogenisation. To assess the need for internal friction in turbulence parameterisations it is important to determine the exact nature of these processes and measure their magnitude in the real ocean.

Acknowledgement

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WOCE/CLIVAR Workshop on Ocean Modelling for Climate Studies

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The workshop was held 10–13 August at the National Center for Atmospheric Research, Boulder, USA, following a joint recommendation by the WOCE SMWG, the WOCE SSG, and the CLIVAR NEG-2 (now the WGCM). It brought together 65 scientists representing the climate modelling, ocean modelling, and observational communities to address the problem of realistically representing the ocean's role in studies of climate change and climate variability, particularly on decadal time scales.

Particular questions the workshop sought to address included:

- What are the key aspects and critical processes that need to be adequately represented in ocean models used to study (decadal-scale) climate variability?
- What are the fundamental limiting factors in the ability of models to capture critical processes? How well do present climate models represent the net-effect of these processes?
- Can we define quantitative data-oriented tests of the models' ability to capture these critical processes?
- What should be done in terms of future modelling efforts to clarify open questions identified during the workshop? In particular, would an "ocean model intercomparison project" (OMIP), or a set of co-ordinated modelling studies be useful?

The meeting was divided (about equally) between plenary sessions with invited speakers asked to address a number of specific questions, and 6 working groups centred around classes of key processes (3-D turbulent mixing; Bottom boundary layer; Mesoscale eddies) and phenomena (Meridional overturning in the Atlantic; Southern Ocean). The recommendations from these working groups will constitute a major part of the final workshop report; they will extend, where feasible, to possible elements of co-ordinated modelling programmes, model diagnostics and datasets that could provide a 'baseline' for model sensitivity and intercomparison studies.

Major issues for the final plenary session, building upon the conclusions and recommendations of the previous plenary and working group deliberations, were the usefulness and requirements for an organised 'Ocean Model Intercomparison Project' (OMIP). Conclusions may be summarised as follows:

- Simulation of large-scale ocean circulation and its response to (thermohaline) forcing variations on decadal and longer time scales is, in contrast to seasonal or interannual (wind-driven) variability, extremely dependent on the representation of interior, often very small-scale processes: e.g., flows through sills and gaps, entrainment in downslope flows, interaction with topographic irregularities. Any organised programme aiming at improving ocean circulation models must take into account these sensitivities and their (sometimes hidden) manifestation in different

numerical model concepts, or choices of horizontal and vertical resolution.

- Understanding the relative roles and impact on model performance of resolution, parameterisations, and numerics will benefit greatly from more co-ordination, transparency and feedback between individual modelling teams. Co-ordinated programmes of experimentation are therefore strongly encouraged. Ocean model intercomparisons should be one possible aspect in such programmes, however, they are meaningful only if embedded in studies investigating sensitivity to the representation of critical physical processes.
- Any meaningful ocean model intercomparison project necessarily has to be of a level that requires participating groups to run models in strictly defined configurations in terms of model domain and forcing. Because of the considerable fraction of resources inevitably going to be tied, OMIPs can be realised only if strictly science-driven: each individual model run has to be of scientific interest, the intercomparison part representing a significant value-adding aspect.
- Because of the foregoing requirements, it cannot be recommended at this stage of ocean model development, to set up a centralised, big 'OMIP'. Instead, ocean modelling groups are strongly encouraged to strive for a closer co-ordination of experimentation by defining and joining in self-organising 'omips'. Since any such effort requires close co-ordination: of model configuration, initial, forcing and boundary conditions, model diagnostics and data sets for evaluation, there will necessarily be limitations on the possible number of partners. (This, however, does not exclude, as previous examples of multi-institutional modelling programmes (e.g., CME, DYNAMO, DAMEE) have shown, the possibility of establishing 'baselines' that may provide useful test-beds for new parameterisation schemes or alternate models.)
- It is recommended that co-ordinated modelling efforts include both high resolution models and models with resolutions comparable to those adopted in present (and next-generation) climate models. Model domains should preferably either be global, or concentrate on Southern Ocean or Atlantic domains. Experience shows that basin-scale models are strongly controlled by the specification of the lateral boundary conditions. However, in some cases this may actually be an advantage, by allowing the avoidance of unwarranted dependence on delicate physics (e.g., interaction with sea ice or ice shelves in polar regions) outside of the domain of interest.
- An issue requiring special attention is integration time, i.e., the question of (thermohaline) equilibrium vs. non-equilibrium simulations. While ocean models ultimately, i.e. for being useful in coupled climate studies, need to demonstrate sufficient realism in equilibrium solutions, there are a number of arguments for putting the

focus of intercomparison efforts on the transient behaviour, in particular, if aiming at improved model capabilities for decadal climate studies. An important one is that equilibrium behaviour tends to be controlled by model factors disjunct from those governing transient behaviour on decadal time scales, exemplified especially in the different impact of adiabatic and diabatic processes. It is recommended that ocean model intercomparison and sensitivity studies extend to an analysis of the transient behaviour to prescribed forcing anomalies, including the question on the relative merit of high-resolution vs. coarse-resolution.

- Horizontal resolution anticipated for the ocean component in the next-generation of climate models (5 to 10 years from now) is on the order of 1–2°. Further decrease in grid size may be possible, but has to be balanced against needs for long coupled runs, ensembles of experiments, and sensitivity runs. Hence it remains important to improve on parameterisations for eddies, narrow passages and boundary currents, and to test the effect of not resolving these and other potentially important phenomena. It is recommended to study the relevance for climate change studies of

ocean model resolution with some carefully-planned eddy-permitting coupled experiments. At least two of such studies are being planned already. It is important that ocean models are built on the best possible parameterisations of critical physics to avoid effects of resolution by other factors.

- While a centralised big OMIP was considered less useful, there was agreement that some centralised infrastructure supporting individual model development and intercomparison efforts would be necessary, in particular, to avoid wasting resources by duplication of efforts concerning construction of data sets for model forcing, initialisation, and testing. Establishment of high-quality, standard forcing fields and protocols, analysis tools, and evaluation data sets was described as an important legacy of any ambitious modelling programme; but mechanisms for facilitating better communication and co-ordination between different efforts need to be considered.

Workshop Committee, responsible for the science and/or local organisation: C. Böning (chair), M. England, D. Haidvogel, L. Talley, E. Tziperman, D. Webb, J. Willebrand, R. Wood, C. Wunsch, P. Gent, W. Large (local organisation).

Southern Ocean: JGR Oceans Special Section – Call for papers

It was agreed at the WOCE Southern Ocean Workshop in Hobart that there should be a Special Section of JGR Oceans for WOCE results from the Southern Ocean. This will follow the Pacific Special Section published in June 1998 and the South Atlantic Section due in 1999.

Plans for the Southern Ocean Section are now developing. Guest Editor will be Brian King, with Mark Warner and David Webb as Associate Guest Editors.

Papers will be welcomed on all aspects of WOCE related to the Southern Ocean. Contributions are not restricted to scientists who attended the Hobart workshop, or to results presented there. In addition to results from individual research projects, contributions from collaborative efforts stimulated by the workshop will be

especially welcome.

Normal JGR procedures for submission and review apply. Papers should be submitted to L. Rothstein, and marked 'Southern Ocean Special Section'. Do not submit manuscripts to the Guest Editors. Final decisions on acceptance rest with the Journal Editor, advised by the Guest Editors.

The deadline for submission of manuscripts is 30 April 1999, with a target publication date early the following year.

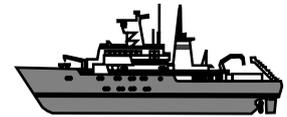
Anyone who intends to submit a paper to this Section is invited to contact the Guest Editor (brian.king@soc.soton.ac.uk) as soon as possible, with details of Title, Authorship and expected submission date.

MEETING TIMETABLE 1999

January 5–6	US WOCE Velocity Workshop	Dallas, USA
January 7–8	US WOCE SSC Meeting	Dallas, USA
February 22–24	Layered Ocean Models Users' Workshop	Miami, USA
February 22–26	WOCE-AIMS Tracer Workshop	Bremen, Germany
March 15–19	WCRP JSC Meeting	Kiel, Germany
April 13–16	WOCE Data Products Committee (DPC-12)	Birkenhead, UK
April 19–23	EGS 24th General Assembly	Den Haag, Netherlands
April 27–30	Extreme and Unexpected Phenomena in the Ocean	Reno, Nevada, USA
May 10–14	CLIVAR SSG-8 Meeting	Southampton, UK
May 31–June 4	AGU Spring Meeting	Boston, USA
July 18–30	IUGG XXII General Assembly	Birmingham, UK
August 23–27	WOCE North Atlantic Workshop	Kiel, Germany
September 13–17	Modelling of Global Climate and Variability Conference	Hamburg, Germany
October	TOPEX/POSEIDON/Jason-1 SWT	Biarritz, France

Offsets of IAPSO Standard Seawater

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Dr Aoyama's article (Issue No. 32) follows a number of publications in the scientific press from various authors on the reliability of Standard Seawater. As detailed in his article, apparent 'offsets' are given for various batches of Standard Seawater when compared with a mean of values from previous batches. Whether made against a single batch or a range of batches, these 'offsets' may be misleading or even incorrect when applied to salinity data retrospectively. One reason for this is that most of the changes which take place in Standard Seawater occur over a period of years so an apparent offset now may not have been present at the time of using the standard. Equally important, is the fact that changes in the reference used may also contribute to this apparent offset. The only unambiguous way of checking such changes is to make measurements against the defined KCl standard, freshly prepared and thus not subject to storage artifacts. The IAPSO Standard Seawater Service is probably the only organisation with detailed experience of such measurements and our findings on changes in the standards with time, as compared with the primary KCl standard have been recently published (Culkin and Ridout, 1998). Table 1 demonstrates that the changes in salinity of the more recent batches (P120–P129) are less than 0.001 in salinity (equivalent to 0.00003 in K_{15}) up to a period of 96 weeks.

Changes in the conductivity of Standard Seawater are thought to occur as a result of microbial activity or interaction between the water and the glass ampoule. Under normal conditions, this is a slow process, therefore 'offsets' may occur some years after the calibration date of the batch (Fig. 1, page 21). Ocean Scientific International endeavour to minimise these offsets by careful control of microbes during the preparation of the standards and by allowing the seawater to equilibrate in the glass ampoules for several months before calibration. Our recommendation throughout the WOCE programme has been to use Standard Seawater not more than 2 years old.

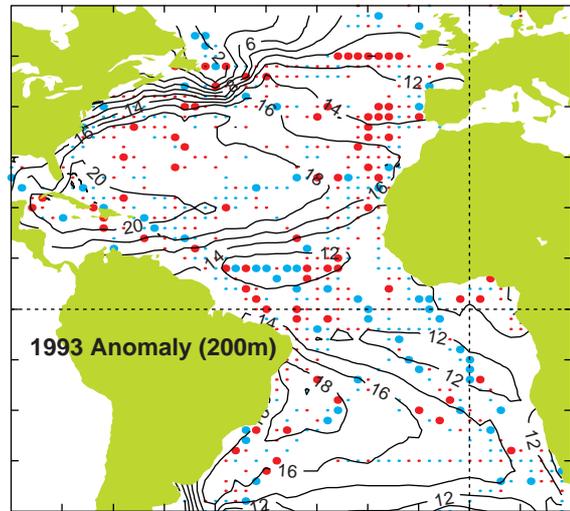
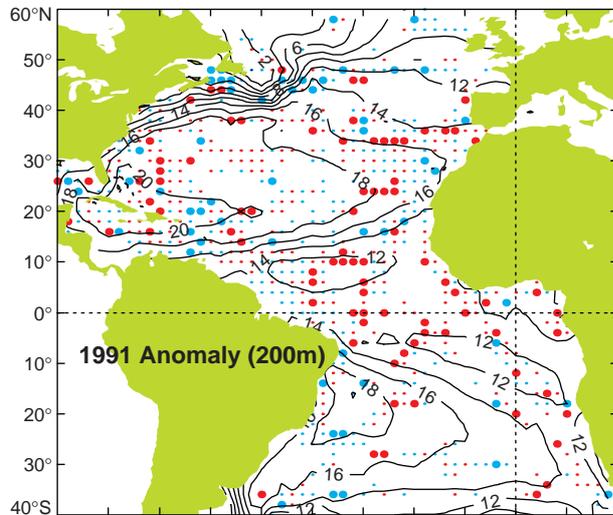
It is opportune at this point, for a reminder that since the Practical Salinity Scale of 1978 (PSS78) was introduced, Practical Salinity is regarded as having no units or dimension. Therefore, it is incorrect to use terms such as PSS, psu, ppt, ‰ after each salinity value.

Table 1. Changes in K_{15} values of IAPSO Standard Seawaters after storage.

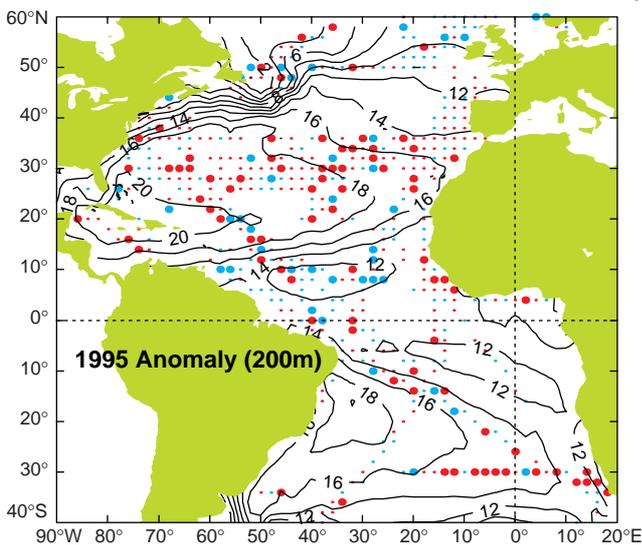
Batch	Date	Age (weeks)	Label K_{15}	New K_{15}	(New-Label) $\times 10^5$	No. of checks
P120	6-May-92	0	0.99985	0.99984	-1	3
	8-Sep-92	18				
	19-Jan-93	37				
	8-May-93	52				
	13-Jan-94	88				
P121	8-Sep-92	0	0.99985	0.99985	0	4
	19-Jan-93	19				
	8-May-93	35				
	13-Jan-94	70				
P122	21-Jan-93	0	0.99991	0.99991	0	6
	8-May-93	15				
	13-Jan-94	51				
	27-Jul-94	79				
	22-Nov-94	96				
P123	10-May-93	0	0.99994	0.99994	0	3
	13-Jan-94	35				
	22-Jul-94	63				
	22-Nov-94	80				
P124	18-Jan-94	0	0.99990	0.99994	1	3
	27-Jul-94	27				
	22-Nov-94	44				
	7-Feb-95	55				
P125	1-Aug-94	0	0.99982	0.99982	0	6
	22-Nov-94	16				
	7-Feb-95	27				
	18-Jul-95	50				
P126	29-Nov-94	0	0.99987	0.99986	-1	4
	7-Feb-95	10				
	18-Jul-95	33				
	21-Nov-95	51				
P127	14-Feb-95	0	0.99990	0.99990	0	2
	18-Jul-95	22				
	21-Nov-95	40				
	19-Mar-96	57				
P128	18-Jul-95	0	0.99986	0.99987	1	7
	21-Nov-95	18				
	19-Mar-96	35				
P129	22-Nov-95	0	0.99996	0.99997	1	5
	19-Mar-96	17				
	16-May-96	25				

Reference

Culkin, F., and P. S. Ridout, 1998: Stability of IAPSO Standard Seawater. *J. Atmos. Oceanic. Tech.*, 15, (4), 1072–1075.

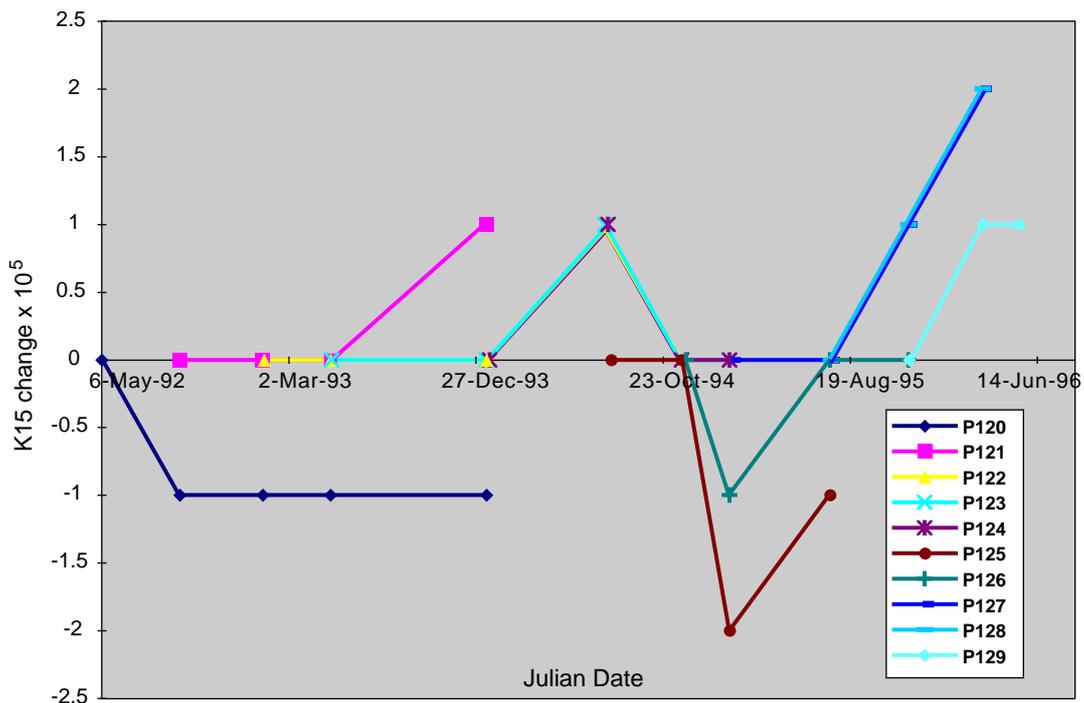


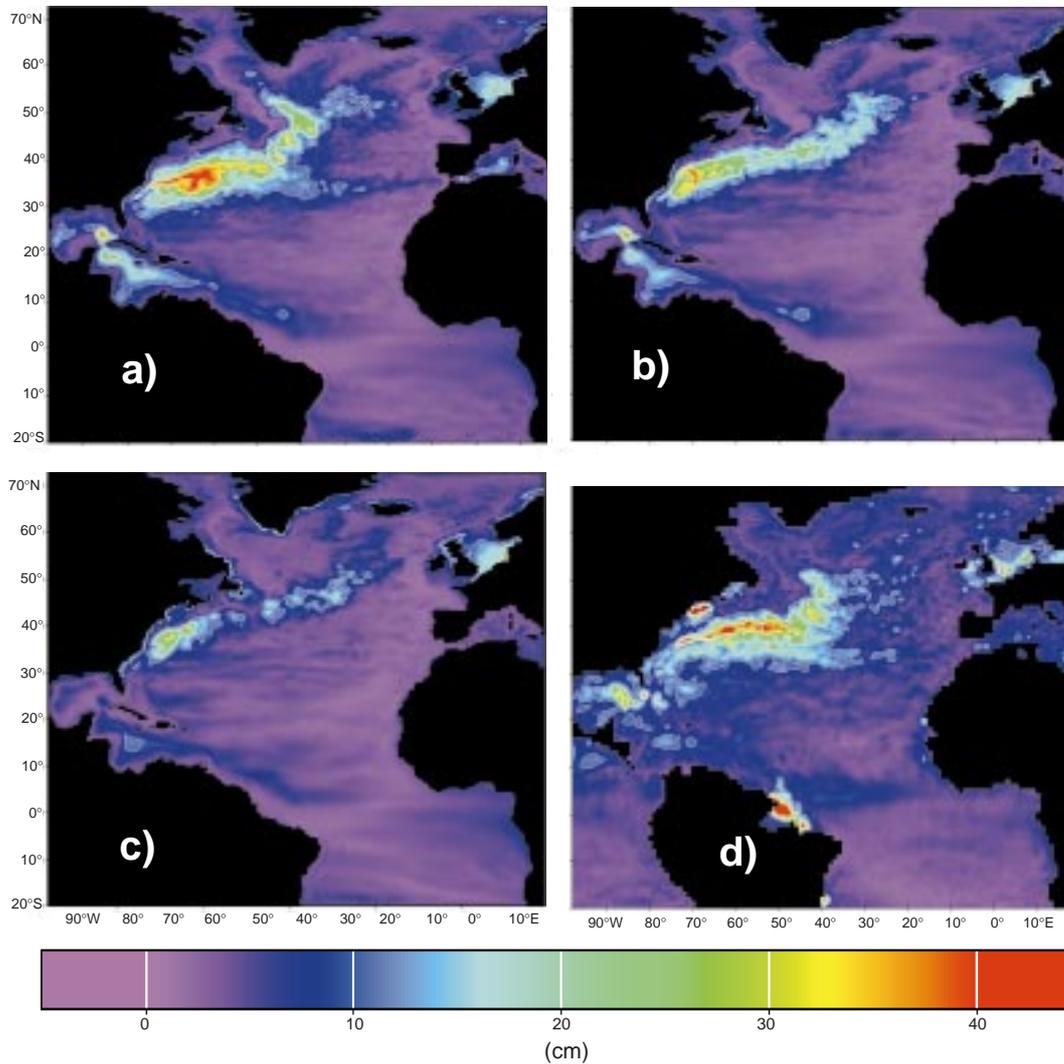
90°W 80° 70° 60° 50° 40° 30° 20° 10° 0° 10° 20°E



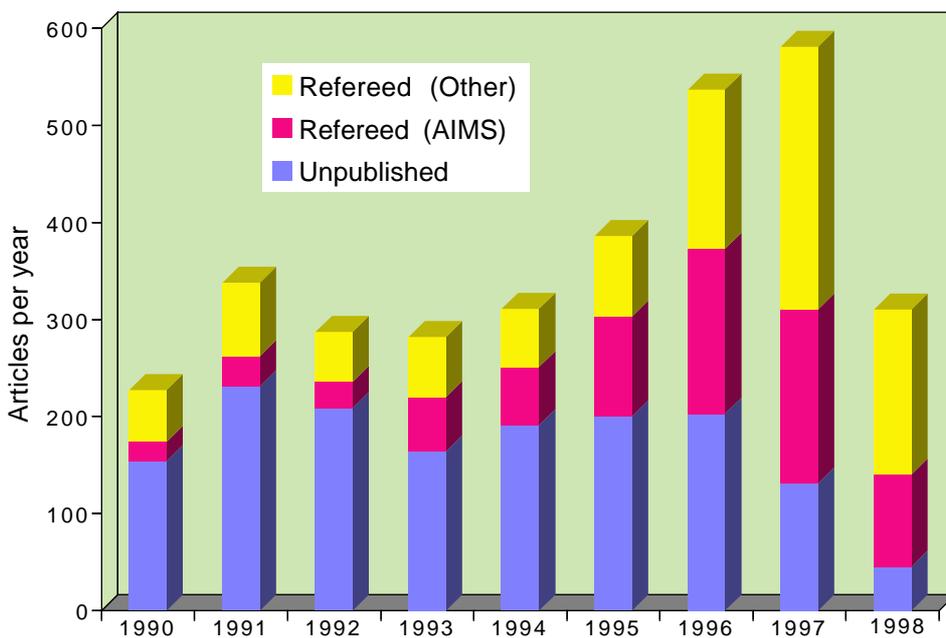
Above and left: Molinari and Festa, page 3, Figure 4. January–February anomalies (200 m) during the WOCE years relative to the long-term climatology given in Fig. 2 and shown as the solid lines in these figures. Grid points with data are indicated by dots. Negative anomalies are marked blue and positive anomalies, red. Anomalies greater or less than one standard deviation from the grid point mean are marked by heavier dots.

Below: Ridout and Culkin, page 20, Figure 1. Changes in K_{15} values of IAPSO Standard Seawaters after storage.

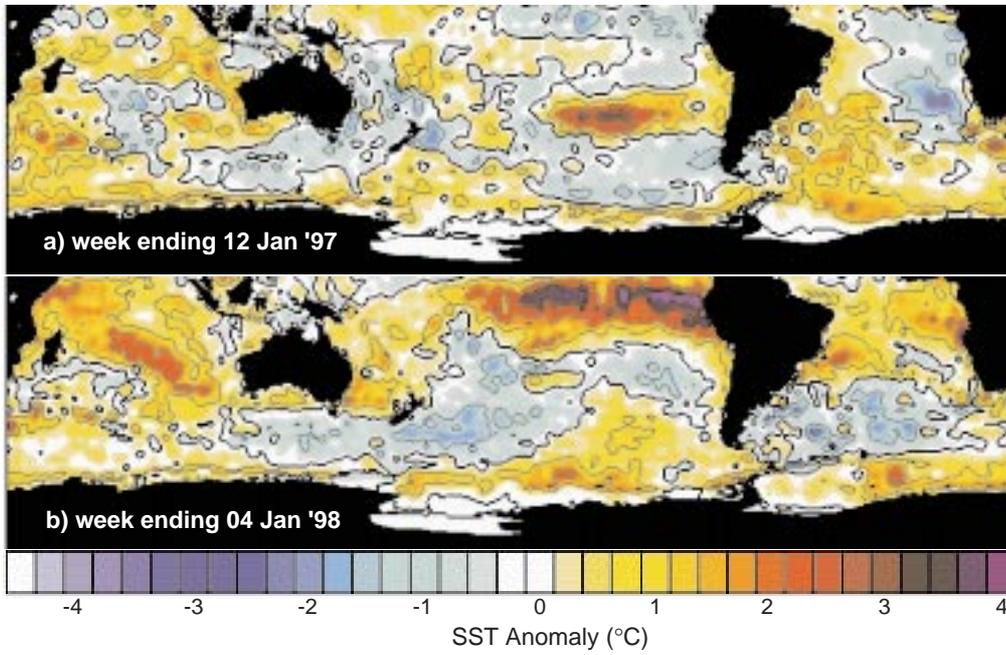




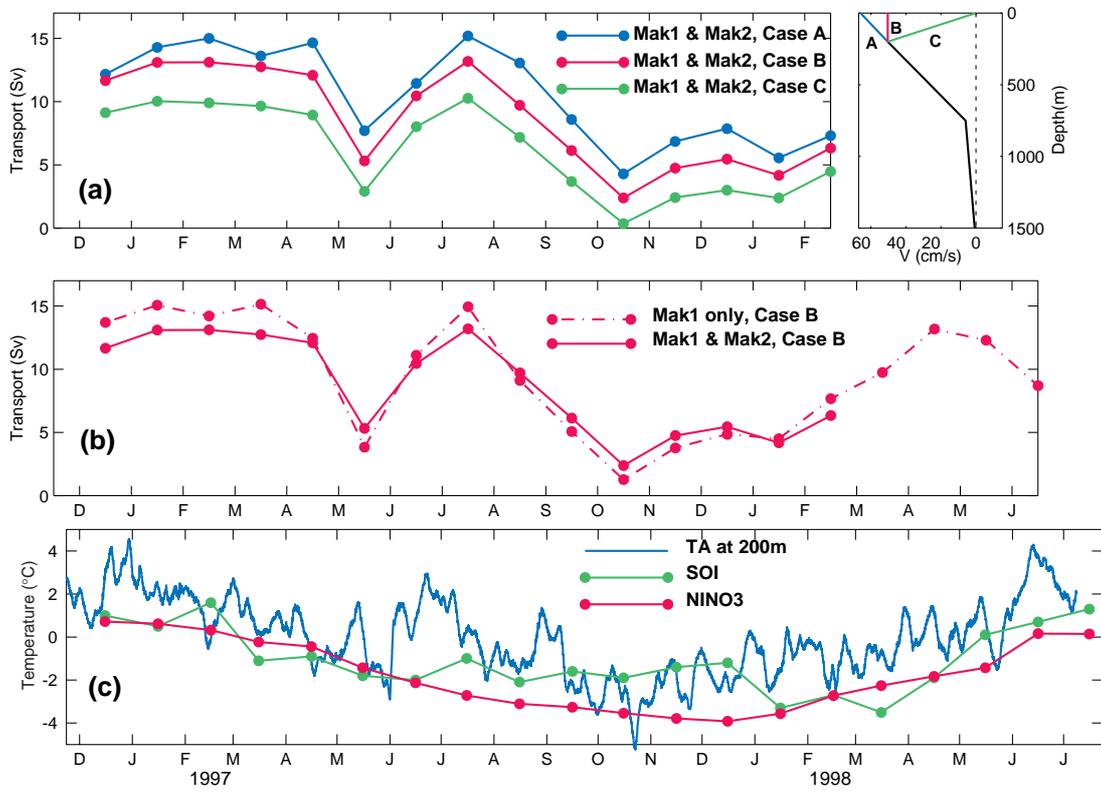
Bryan and Smith, page 12, Figure 1. RMS sea surface height variability of the North Atlantic models at (a) 0.1°, (b) 0.2°, (c), 0.4°, and (d) for a blended ERS-TOPEX/POSEIDON altimetric product (Le Traon and Ogor, 1998).



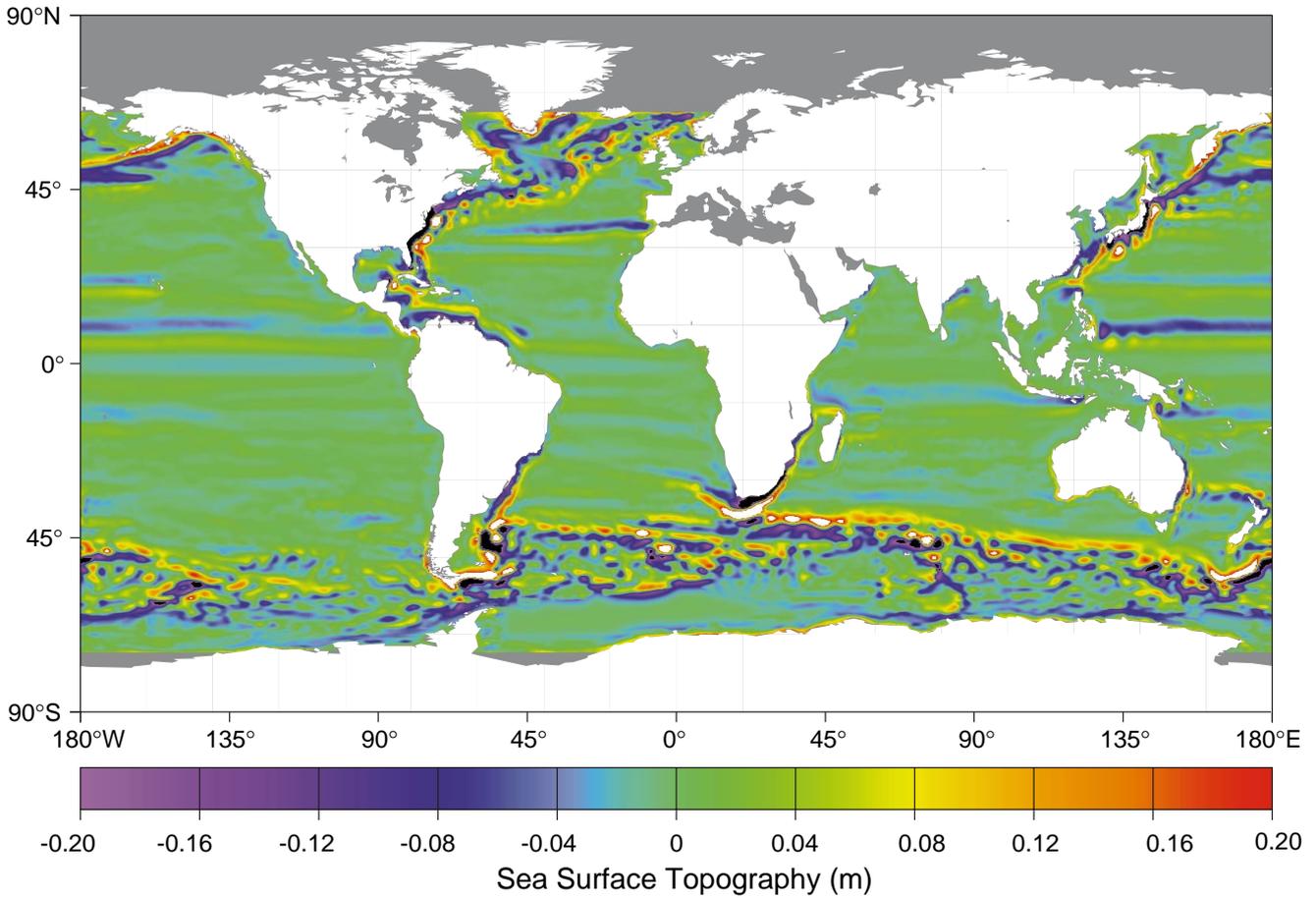
Saunders et al., page 42, Figure 1. Refereed and non-refereed (unpublished) articles in the WOCE bibliography. The 1998 inventory is for 8 months only.



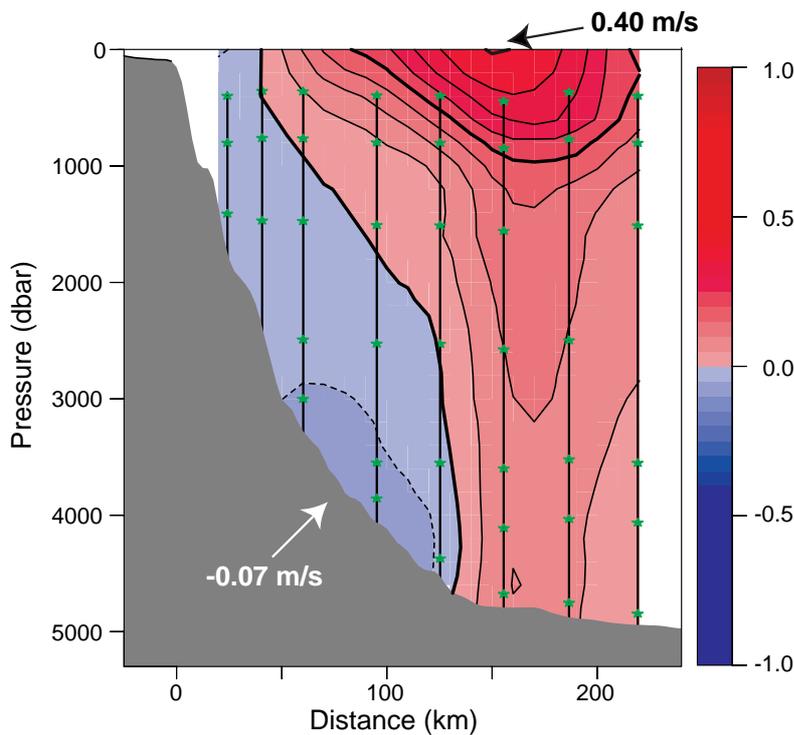
Baines, page 35, Figure 1. SST anomalies for the Southern Hemisphere for the weeks ending (a) 12 January 1997 and (b) 4 January 1998, approximately 12 months apart. The ACW is evident in the Southern Ocean region, and the 12-months delay shows the eastward movement. Summer pictures are best because of the reduced cloud cover in the region. (Pictures by courtesy of the web page of Neville Smith, BMRC.)



Gordon et al., page 30, Figure 4. Total southward or throughflow transport within Makassar Strait (displayed as positive values) for each month (Fig. 4a,b). As the data from the moored ADCP (deployed at 150 m) is still being processed (see text) we use three models for carrying the Aanderaa along channel speeds to the sea surface (insert adjacent to 4a). As mentioned in the text, we favour case B, which on average differs from A and C results by 2.4 Sv, about 25% difference. Transport determined from use of both moorings for the period up to February 1998, agrees closely with the use of only MAK-1 (4b), suggesting that one mooring may be sufficient in monitoring Makassar transport. The temperature recorded by the 200 m instrument of MAK-1 was processed to remove the mean vertical temperature gradient, recorded during the strong semi-diurnal blowover movements. An anomaly of temperature was then calculated, the difference between the temperature change expected from the mean temperature profile from that actually recorded. The temperature anomaly (Fig. 4c) compares favourably to the ENSO indicators of SOI and the SST anomaly at El Niño-3. The thermocline is deeper during La Niña, shallower during El Niño, as also shown in Fig. 2; thermocline depth is correlated with transport magnitude.



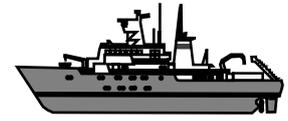
Woodworth et al., page 37, Figure 4. Ocean topography signals with spatial scales of degree 80 (500 km wavelength) and shorter, derived from the mean topography of the 8 year model run of the Semtner-Chervin model (courtesy of Dr R. Tokmakian).



Clarke et al., page 33, Figure 1. Mean Transverse Flow (August 1993 to April 1995) for Current Meter Array ACM6. Distance scale has its origin at the 200 m isobath on the Tail of the Grand Banks. Contour intervals are 0.05 m/s, red shading represents flow to the northeast, blue to the southwest.

An Example of Ageing in IAPSO Standard Seawater

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It is normal to regard the WOCE requirement for salinity accuracy as 0.001, but the relevant part of the operations manual (WHPO, 1994) states that 0.002 salinity accuracy can be expected given individual samples analysed to an accuracy of 0.001 and IAPSO standard seawater (SSW) salinity known to 0.001. If the salinity of SSW is known 'nearly exactly', then the absolute accuracy of sample measurements may approach 0.001. This issue becomes important as scientists seek to identify climate change in the deepest and most slowly varying waters of the world. There is also the more immediate issue of identifying and correcting salinity mismatches between different cruises which visit the same place at nearly the same time. It is known that some early batches of IAPSO SSW suffered from inaccurate salinity (Mantyla, 1980, 1987, 1994), some by over ± 0.003 , and there is continuing interest in the attempt to verify or determine the salinity of SSW as used routinely at sea.

This note was provoked by Aoyama et al. (1998; A98 below) who attempt to update (to close to the present) the history of SSW 'corrections'. We demonstrate below that simple corrections for modern SSW will not do. Modern SSW is produced correct (actual salinity equal to label salinity) at source, as demonstrated by Culkun and Ridout (1998; CR98 below). They stress that the salinity of SSW can evolve over months or years while in the ampoule and suggest that reactions between seawater and glass, or microbial activity, are the cause. We provide below an

example of this from a recent cruise, Discovery 230 (Bacon, 1998).

On this cruise, run in August–September 1997, we used SSW batches P128 (the oldest, 25 months old), P130 (17 months old), P131 (9 months old) and the newest, P132, at 4 months old. We used batches in groups on this cruise, so we calculate mean differences from steps using 6 measurements from each side of the step, except for the final comparison set, which were all interleaved; the salinometer standardisation history is shown in Fig. 1. The use of steps around batch changeovers avoids biasing the inter-batch differences with the long-term drift in the salinometer response. In all cases of mean salinity difference calculation, the standard deviation of the difference about the mean difference is 0.0004 or better; therefore the standard error of the mean difference is 0.0002 or better for 6 samples. There is one exception which we will mention below.

The notation which follows is: C is measured deviation from label salinity (in terms of salinity differences) having first eliminated the measured salinity difference due to label salinity; subscript is batch number; superscript is comparison set or step number from Fig. 1. There are three batch changes (numbers 1–3) and three independent differences within the final comparison set (number 4). The results are as follows: $C_{130}^1 - C_{128}^1 = -0.0008$, $C_{131}^2 - C_{130}^2 = 0.0000$, $C_{132}^3 - C_{131}^3 = -0.0008$; and $C_{130}^4 - C_{128}^4 = -0.0011$, $C_{131}^4 - C_{130}^4 = -0.0005$, $C_{132}^4 - C_{131}^4 = -0.0001$. We assume that P132 was at its label salinity since it was only 4 months old at time of use, so that P131 appears 0.0001 or 0.0008 salty; P130 appears 0.0006 or 0.0008 salty; and P128 appears 0.0016 or 0.0017 salty. These difference calculations are generally very consistent, but those for P131 were rather more scattered than for the others (s.d. of differences 0.0009), which at the time of the cruise led to our discontinuing the use of this batch. Now CR98 offer no SSW measurements beyond batch P129, so we can only look there at P128 measurements when the batch was 9 months old, at which time it had grown more salty by 0.00002 in K15, equivalent to 0.0008 in salinity. Our measurements were made a further 17 months after that.

We have an independent check on our results. P128 was used as standard while the ship was in the eastern North Atlantic on $41^{\circ}30'N$, so we can compare the bottle salinities with the salinity predicted by Mantyla's (1994) latitude-dependent θ/S relationship, after the manner of Saunders' (1986) use of the deep ocean as a natural calibration tank. The results are shown in Fig. 2, which incorporates data from stations 20–36 (Bacon, 1998). The mean salinity offset (samples minus Mantyla line) for $2.1 < \theta < 2.4^{\circ}C$ is

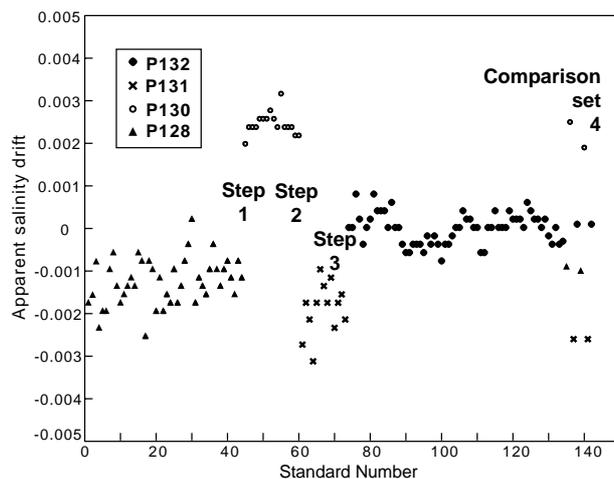


Figure 1. Salinity standardisation history from Discovery Cruise 230. Standard seawater batches are identified by the key at the top left. Apparent salinity drift is referenced to an arbitrary zero. Steps number the batch changes described in text, the comparisons made within the final group of measurements (set 4) are numbered in the source sequence.

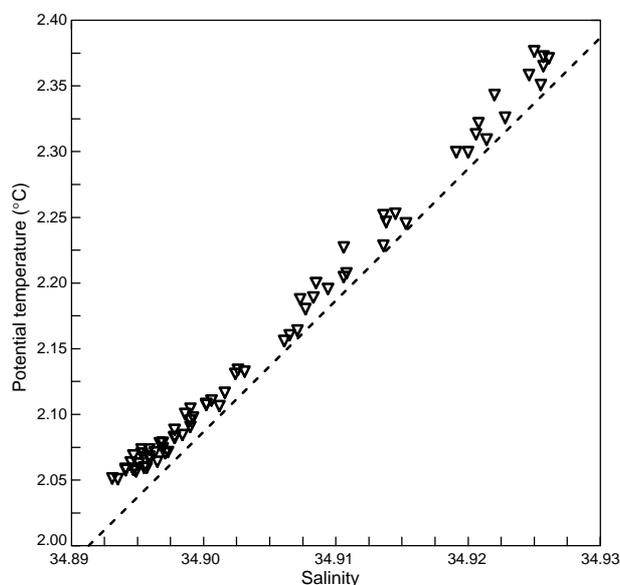


Figure 2. Deep θ/S bottle data from Discovery Cruise 230 (uncorrected). Stations are in the western North Atlantic along $41^{\circ}30'N$. The line shows that the θ/S relationship predicted by Mantyla's (1994) formulation for that latitude.

-0.0019 (s.d. 0.0009), which is nearly identical to our inferences above (a too-salty standard results in overcorrected measurements which thereby appear fresh).

Now there is an important difference between A98's 'offsets' and our age-dependent salinity changes. They claim P128 to be 'correct' (0.0001 different from label salinity) whereas we show with some confidence that at the time of use on Discovery 230, it was 0.0015–0.0020 salty. It is likely that A98's measurement of P128 was made close to the time of production, whereas ours was made over 2 years later. We caution against the uncritical use of such offsets, which have already been applied in a study by Gouretski and Jancke (1998) of WOCE section crossings. They find that application of the 'offsets' of A98 results in a decrease of mean absolute section-to-section salinity difference from 0.0023 to 0.0019, wherein individual differences are decreased in 32 cases but increased in 20 cases. We believe that the cases of worsening are caused at least partly by the application of 'corrections' inappropriate to the age of the SSW batch used on the relevant section cruises.

We are preparing a more complete study of SSW as used in UK WOCE cruises (Bacon, Snaith and Yelland, 1998); the most important message of both this and the present note is that the crucial factor in assessing inter-cruise salinity differences from the point of view of SSW is the age of the SSW at the time of use. We believe SSW to be of 'correct' (label) salinity at the time of production and within specification within 18 months of production (i.e. not more than 0.001 different from label salinity). Batches used at greater ages than this should be regarded with circumspection; the type of analysis demonstrated here is one means of determining any salinity change.

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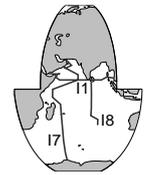
DIU Requests Information on Indo/Pacific Throughflow

While the determination of the exchange of mass, heat and salt between the Pacific and Indian Oceans is important to achieving WOCE objectives, direct measurements of these exchanges were not given as high priority as other WOCE measurements. Fortunately, there were other non-WOCE measurements in the Throughflow region during 1990–97, the WOCE observational period. These measurements, if combined with those organised within WOCE, could provide a more complete basic dataset for the analysis, interpretation, modelling and synthesis being planned during WOCE's final phase and for future ocean circulation studies.

The first step is to put together a complete list of Throughflow datasets obtained during WOCE. We are starting, of course, with WOCE measurements and datasets that have been identified in various newsletters (see, for example, *International WOCE Newsletters* 22, 24 and 27) and other journals but we are certain that this will only provide information on a fraction of the data that exist. We would like to request that anyone having or knowing of the existence of Throughflow data, inform the WOCE Data Information Unit, specifically James Crease or Bert Thompson (woce.diu@diu.cms.udel.edu). They need: PI, type of data, where, when, data status and availability, publication references and so on.

Inclusion of these data in the WOCE datasets now being processed and issued (first CDs issued May 1998, second set early 2000) may be proposed but would only be done with PI agreement.

Comparison of Bottle Salinity and Bottle Oxygen Values from WHP Repeat Lines I7N, I1W, I8N and I8S



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In early 1995, the NOAA ship Malcolm Baldrige began a global circumnavigation that included occupation of WOCE Hydrographic Programme (WHP) track lines, I7N, I1W, I8N and I8S (Fig. 1). These cruises were designed to repeat occupations of WOCE “one time” transects by the WHOI ship RV Knorr (Table 1). The goal of the repeat cruises was to obtain data during opposing monsoon seasons in order to define the large seasonal variability of the Indian Ocean due to the reversal of seasonal monsoon winds. In total, 310 CTD stations were completed by the Baldrige. Hydrographic measurements obtained during these cruises were temperature, salinity, oxygen, current velocity and a suite of chemistry and atmospheric variables. This report will

compare the Baldrige salinity and oxygen bottle data to the Knorr salinity and oxygen bottle data taken along the same track lines to determine the suitability of these observations for comparative water mass studies.

Hydrographic data onboard the Baldrige was obtained with a Seabird 911 plus CTD, deck unit, and rosette pylon. Twenty-four 10 litre PVC Niskin bottles were mounted on a rosette frame, along with the CTD, pinger, lowered acoustic doppler current profiler (LADCP), and LADCP external battery pack (Ffield et al., 1998). Water samples were taken after each cast and analysed for salinity, oxygen, and other chemical and atmospheric measurements.

A multi-step procedure was used to determine differences between the

Baldrige and Knorr salinity and oxygen bottle data. It was suggested to display the differences between these data sets as a function of potential temperature. After analysing the data in this way, statistical analysis showed there was no substantial difference between salinity/oxygen versus pressure and salinity/oxygen versus potential temperature. Therefore, it was decided to interpret the data with respect to pressure. Baldrige and Knorr positions were compared and any stations that fell within 0.10° were considered a match. Preliminary comparisons of matched stations indicated that comparisons were best accomplished for portions of the water column with small gradients in the vertical component. These depths were different for each section. The selected depths are given in Table 2. Erroneous bottle values were identified and eliminated from the

Table 1. Completed repeat WHP track lines with ship that occupied the line, date of the cruise and chief scientist.

WHP line	Ship	Dates	Chief Scientist
I7N	Malcolm Baldrige RV Knorr	21 March–22 April 1995 15 July–24 Aug 1995	Amy Ffield Donald Olson
I1W	Malcolm Baldrige RV Knorr	31 May–30 June 1995 29 Aug–16 Oct 1995	Bob Molinari John Morrison
I8N/S	Malcolm Baldrige RV Knorr RV Knorr	22 Sept–25 Oct 1995 10 March–15 April 1995 1 Dec 1995–19 Jan 1996	Bob Molinari (I8N/S) Lynne Talley (I8N) Mike McCartney (I8S)

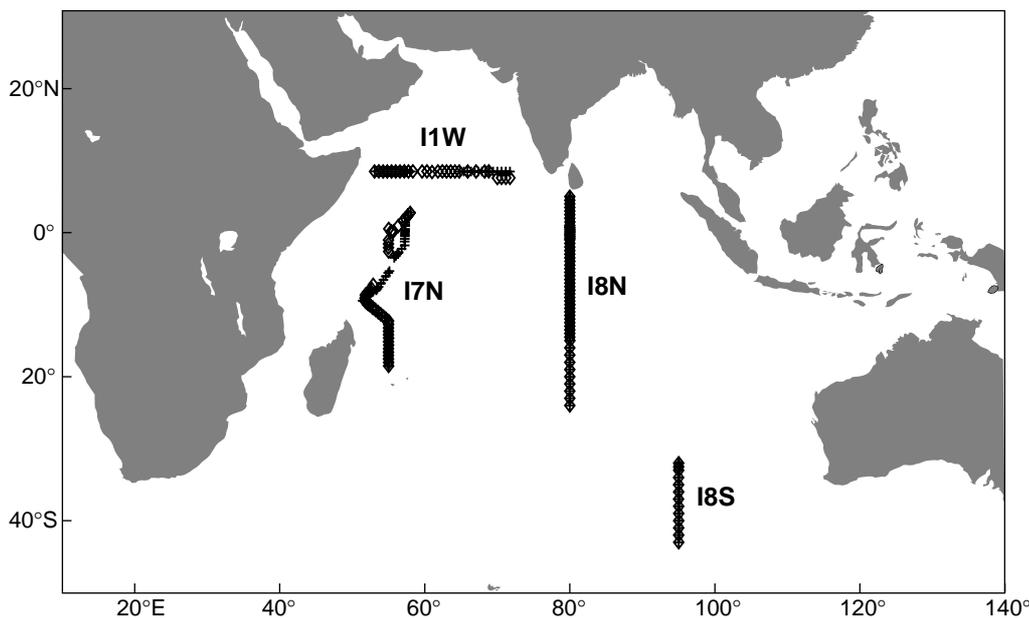


Figure 1. WOCE Hydrographic Programme (WHP) track lines completed during the 1995 field season of the NOAA ship Malcolm Baldrige (diamonds) and the WHOI ship RV Knorr (crosses).

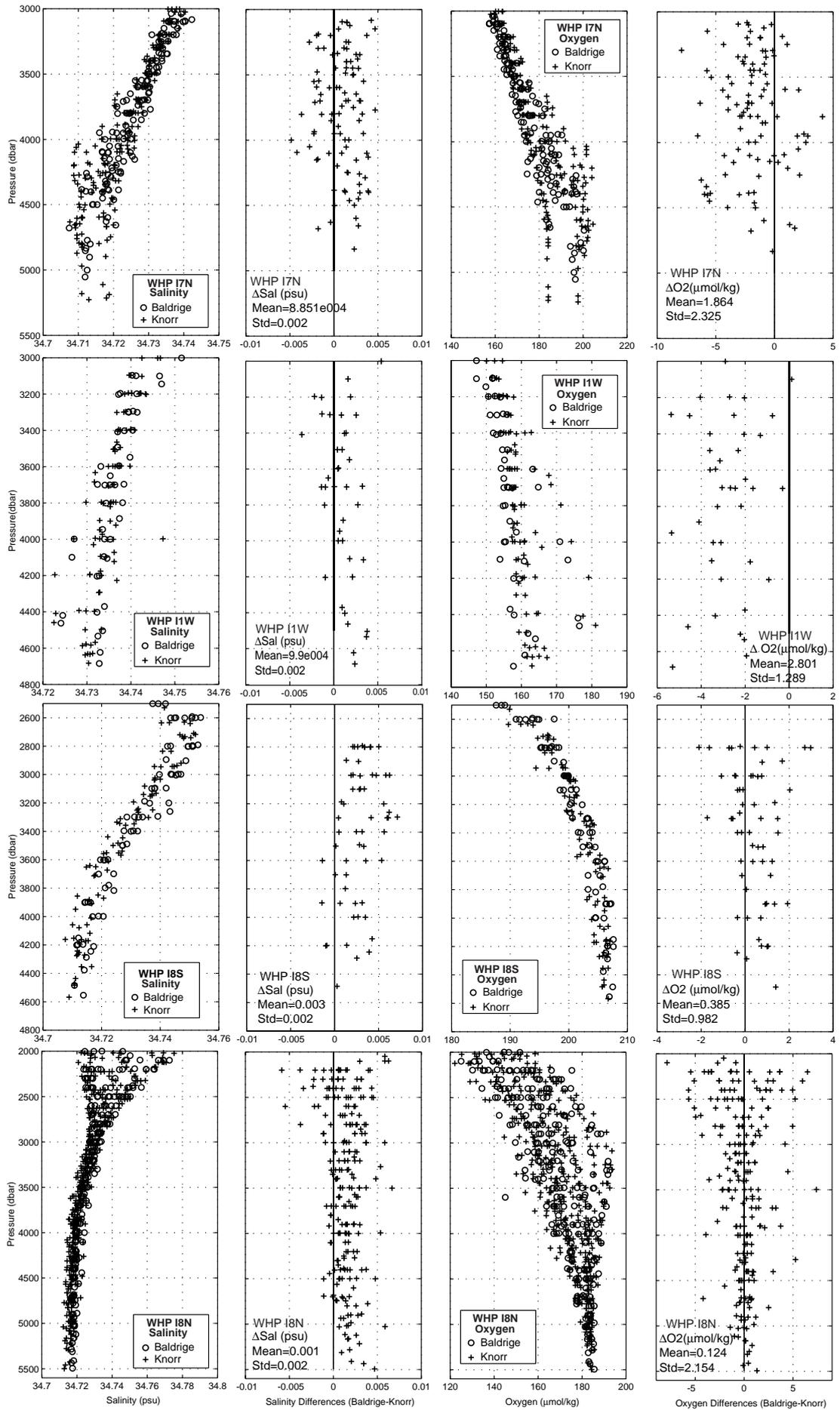


Table 2. Mean and standard deviations of bottle salinity and bottle oxygen differences for all three repeat WHP lines completed by the Baldrige and the Knorr. Depth ranges and number of points used to determine mean and standard deviations are also shown.

WHP line	Depth range	Number of points used	Mean salinity differences	Salinity differences std. deviation	Mean oxygen differences ($\mu\text{mol/kg}$)	Oxygen differences std. deviation ($\mu\text{mol/kg}$)
I7N	>3000 m	641	0.00089	0.0021	-1.864	2.325
I1W	>3000 m	328	0.00099	0.0018	-2.801	1.289
I8S	>2500 m	163	0.003	0.0020	0.385	0.982
I8N	>2000 m	773	0.001	0.0018	-0.124	2.154

analysis. Finally, Knorr salinity and oxygen data were interpolated to Baldrige pressures, and differences were computed. Composite S and O diagrams are plotted in Fig. 2. Table 2 gives the mean and standard deviation for bottle salinity and oxygen differences for all four repeat WHP lines completed by the Baldrige and the Knorr during the 1995 field season.

According to WOCE specifications the repeat hydrography sections must meet certain data quality standards. For salinity, WOCE requires 0.002 accuracy. For oxygen, reproducibility <1%; precision 0.1% (Joyce and Corry, 1994). The salinity and oxygen values described in Table 2 fall well within the WOCE data quality standard requirements. One possible cause for salinity differences may be due to the use of different Standard Sea Water batches (Table 3). The Knorr used batches 124 through 126 while the Baldrige used batch 125 for the entire field season. The differences in the oxygen values may be due to the use of unique in situ oxygen sample analysis systems.

The close comparisons of the Baldrige bottle data versus the Knorr bottle data indicate that the data are suitable for studies of changes in the water mass properties between occupations of the various stations. These close comparisons also indicate that deep water values did not change over the half-year time scale.

Table 3. Standard Sea Water batches used during the 1995 field season of the NOAA ship Malcolm Baldrige and the WHOI ship RV Knorr with salinities (psu) and standard sea water batch differences.

WHP line	Ship	Standard Sea Water batch used	Standard Sea Water batch salinity	Standard Sea Water batch differences
I7N	Baldrige Knorr	125	34.993	-0.002
		126	34.995	
I1W	Baldrige Knorr	125	34.993	-0.003
		123	34.998	-0.005
		124	34.996	
I8S	Baldrige Knorr	125	34.993	-0.003
		124	34.996	
I8N	Baldrige Knorr	125	34.993	-0.002
		126	34.995	

Acknowledgements

The extensive efforts of the officers and crew of the NOAA ship Malcolm Baldrige and the Woods Hole ship RV Knorr are gratefully acknowledged. Contributions by scientific and technical personnel Bob Molinari, Amy Ffield, Lynne Talley, Don Olson, Jay Harris, Mike McCartney, John Morrison, Ryan H. Smith, and W. Douglas Wilson are greatly appreciated.

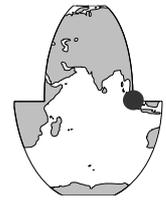
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Facing page: Figure 2. Profiles for all four repeat WHP lines with composite ΔS (salinity) and ΔO (oxygen) differences. Differences were calculated by subtracting Baldrige values from interpolated Knorr values. Baldrige values are circles and interpolated Knorr values are crosses.

Makassar Strait Transport: Preliminary Arlindo Results from MAK-1 and MAK-2

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Makassar Strait is the primary pathway of the Pacific to Indian Ocean transport referred to as the Indonesian Throughflow. The transport through Makassar Strait was measured as part of the Indonesian-USA Arlindo programme, at two moorings deployed within the Labani Channel, a deep (2000 m) constriction (45 km) near 3°S (Fig. 1). Both moorings were operative from December 1996 to February 1998, a 1.4 year time series, when the MAK-2 mooring was released and recovered; the MAK-1 mooring was recovered in early July providing a 1.7 year record. The MAK moorings were deployed during a weak La Niña phase. An El Niño condition began in March 1997, becoming extreme during 1997 summer and fall, relaxing in early 1998. Arlindo data will be the subject of much study by the Arlindo research team†, but because of WOCE interest in Indonesian Throughflow we offer this preview of Makassar transport based on the Aanderaa current meters (current, temperature, pressure) and temperature-pressure pods of MAK-1 and MAK-2.

The Indonesian maritime continent with its complex network of passages and basins connecting the Pacific and Indian Oceans inhibits free communication between the Pacific and Indian Oceans. Schneider (1998) using a couple model, shows that the presence of the Pacific to Indian interocean transfer shifts the warmest SST and associated

atmospheric convective region towards the west, relative to a no throughflow condition. Webster et al. (1998) state: the Indonesian Throughflow heat flux "...is comparable to the net surface flux over the northern Indian Ocean and a substantial fraction of the heat flux into the western Pacific warm pool...it would appear that the Throughflow is an integral part of the heat balances of both the Pacific and Indian Oceans."

Observations indicate that the Throughflow is composed mostly of North Pacific thermocline and intermediate water flowing through Makassar Strait (Gordon and Fine, 1996), which then passes into the Indian Ocean through the passages of Lesser Sunda Islands. East of Sulawesi South Pacific water infiltrates the lower thermocline and dominates the deeper layers, including the Lifamatola Passage overflow into the deep Banda Sea (Van Aken et al., 1988; Gordon and Fine, 1996; Hautala et al., 1996), but it is unlikely that the eastern channels carry total more than 3 Sv.

Indonesian Throughflow estimates based on observations, models and conjecture range from near zero to 30 Sv (Godfrey, 1996). Measurements in the Lombok Strait in 1985 (Murray and Arief, 1988; a near zero SOI value) indicate an average transport of 1.7 Sv. Molcard et al. (1996) as part of the French-Indonesian programme JADE, find a mean transport to the Indian Ocean of 4.3 Sv

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† The US scientists involved in Arlindo Circulation are: Arnold L. Gordon (US chief scientist for Arlindo; CTD) and Amy Ffield (temperature-pressure pods), Lamont-Doherty Earth Observatory of Columbia University; Silvia Garzoli (PIES), NOAA Atlantic Oceanography and Meteorology Laboratory; Dale Pillsbury (current meters) Oregon State University; Rana Fine (CFC) University of Miami; Chester Koblinsky (satellite data), Goddard Space Flight Center. The Arlindo Indonesian Team include: R. Dwi Susanto (BPPT and LDEO); A. G. Ilahude (LIPI and LDEO); from BPPT: Muhamad Irfan, Handoko Manoto, Fadli Syamsudin, Djoko Hartoyo, Bambang Herunadi; and from LIPI: Muhamad Hasanudin.

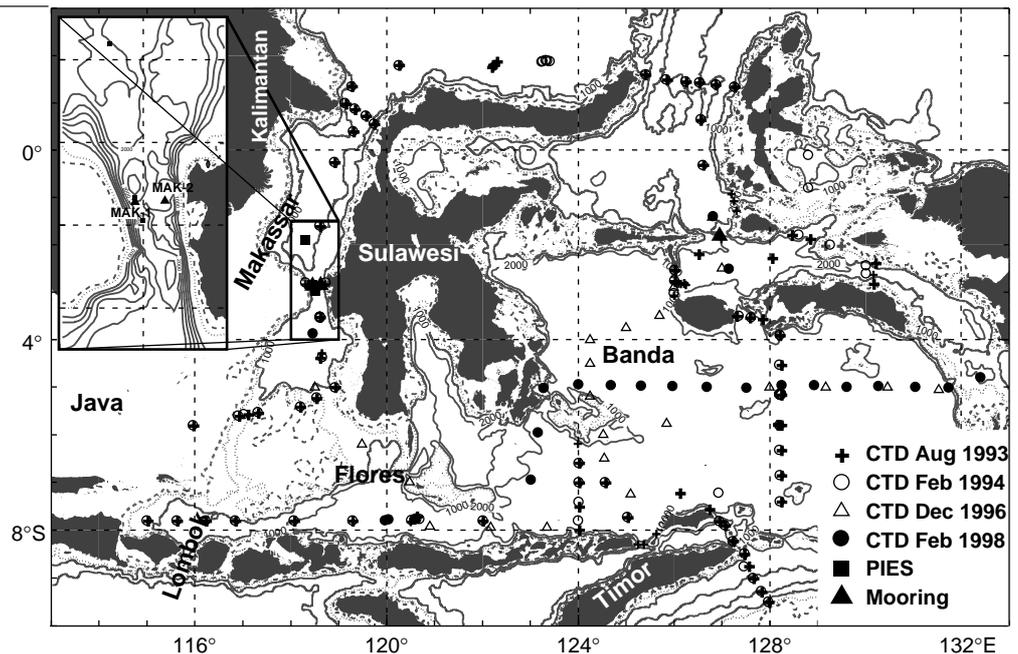


Figure 1. Distribution of CTD stations and time series moorings obtained by the Arlindo programme. The position of the current meter and temperature-pressure pod moorings in Makassar Strait, MAK-1 (2°52' S, 118°27' E) and MAK-2 (2°51' S; 118°38' E) the subject of this note, are shown in the insert, as are the Pressure, Inverted Echo Sounder sensors (PIES).

between the sea surface and 1250 m in the Timor Passage from March 1992 to April 1993 (an El Niño period). With the Lombok values the JADE results suggest 6 Sv transport through the Lesser Sunda Islands. An annual mean Throughflow of 5 Sv is estimated from XBT data for the upper 400 m between Java and Australia for the period 1983 to 1989 (Meyers, 1996). T. H. Aung presented the results of the 1993–94 (El Niño period) ASEAN current meter array in the Makassar Strait at a June 1995 meeting in Lombok. The three ASEAN moorings were deployed in the wide northern entrance to Makassar Strait. Most of the current meters were below 400 m

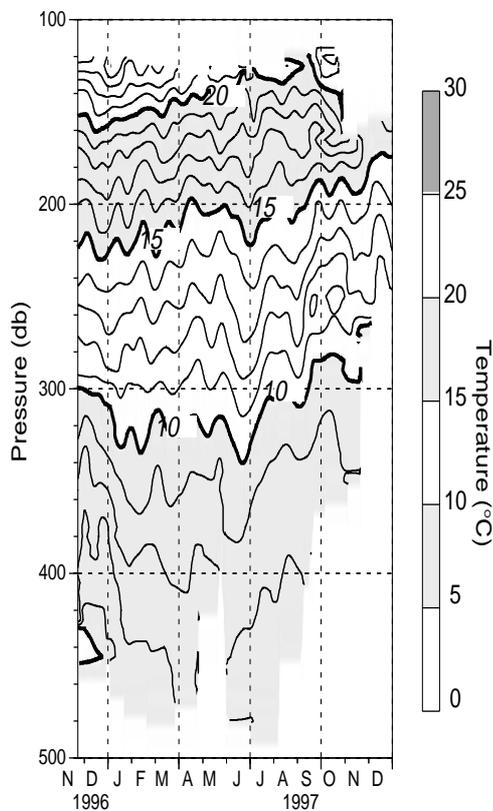


Figure 2. Temperature time series section constructed from 7 temperature-pressure pods on the MAK-1 mooring distributed between 110 m and 290 m. Due to a ~300 m mooring blowover by strong semi-diurnal tidal currents, a continuous pressure-time section was obtained. Each temperature-pressure pod effectively sampled 4 vertical profiles per day. Instrument failures reduce the time series to less than 1 year, rather than the full mooring deployment. The data have been time smoothed by a 35 day gaussian filter. The 15°C isotherm shallows 50 db between November 1997 and January 1998, with the 200 db temperature cooling 2°C over the same time interval.

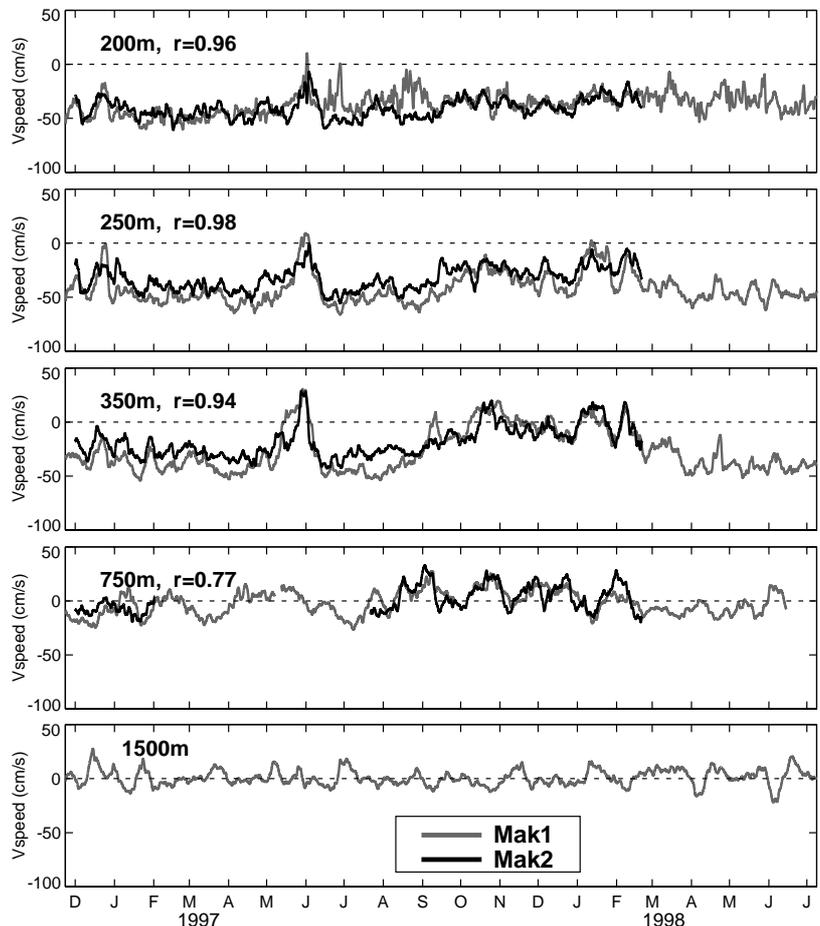


Figure 3. The low pass (2 day) filtered along channel (orientation of 170°) speed recorded at each Aanderaa current meter of MAK-1 and MAK-2. Negative values denote flow towards the south, the direction of the interocean Throughflow. The correlation (r) at similar depths between the two moorings is quite high for the shallower three levels. The lower r value at 750 m stems from the failure of the MAK-2 instrument at that level for nearly a 6 month period. Only MAK-1 had an instrument at 1500 m. The effects of the ENSO (see Fig. 4c for the SOI and El Niño indices) phasing may be seen in the Throughflow speeds: higher southward speeds occur from December 1996 to August 1997, with lower values in the September 1997 to February 1998 period, before increased speeds in the later part of the MAK-1 record. From mid-May to early June 1997 a marked relaxation in the Throughflow speeds is recorded; this event may reflect remote forcing from the Pacific or Indian Oceans. At the 750 m and 1500 m depths, which are below the 600 m sill depth separating the Makassar Strait from the Flores Sea to the south, reveal nearly zero mean flow, but display strong monthly oscillations.

with one instrument at 275 m, thus missing the main thermocline, making estimation of transport difficult, but Aung states that the Makassar transport may be as large as 11 Sv. Potemra et al. (1997) model study and inspection of TOPEX/POSEIDON data find a summer maximum of 11 Sv, and a winter minimum of 4 Sv, with a 7.4 Sv 9-year mean. Gordon et al. (1997) find on average 9 Sv of Indonesian Throughflow water advected westward within the Indian Ocean.

The preliminary findings of the Arlindo Makassar MAK-1 and MAK-2 data are presented in Figs. 2 to 4, some key points each of which will be explored in detail by the Arlindo team, are:

1. The Makassar thermocline depth and transport reflect the phases of ENSO, with an ambiguous seasonal cycle: deeper thermocline, larger

throughflow during La Niña; shoal thermocline, with reduced transport during El Niño. Additionally, during the El Niño months December 1997 to February 1998 the transport average is 5 Sv, while during the La Niña months of December 1996 to February 1997 the average is 12 Sv, a 2.5 fold difference.

2. Along channel flow exhibits much activity at frequencies above seasonal. A special event occurs in May and June 1997 when a marked relaxation of the Throughflow transport is recorded. Candidates responsible are: Pacific Ocean Rossby waves, Indian Ocean coastal Kelvin waves, local atmosphere and dynamics internal to the Indonesian Seas.
3. The Makassar Strait 1997 twelve month average throughflow is 9.3 Sv. This assumes that the flow above the shallowest Aanderaa equals the flow at that current meter (case B, Fig. 4, page 23). Other models for the surface flow yield 1997 transport average of 6.7 Sv (zero surface flow, case C, Fig. 4) to 11.3 Sv (thermocline shear is extrapolated to the sea surface, case A, Fig. 4). How to handle the water flow above the shallowest Aanderaa current meter is an important issue, not just for the mass transport but also for the interocean heat and freshwater flux and for monitoring array design. We will have a firmer idea of the surface layer flow when the moored ADCP data are processed. The MAK-2 ADCP has a record from 1 December 1996 to 9 March 1997 before it flooded; the MAK-1 ADCP data record will be processed later this year. The preliminary MAK-2 ADCP data show a maximum of along channel flow at 110 m, with near zero surface flow. The hull ADCP of the Baruna Jaya IV, the Indonesian research vessel used in the November/December 1996 deployment and February 1998 recovery of the MAK moorings reveal similar reduction of along channel speed in the surface layer. The MAK-1 monthly along channel speeds displays higher southward values at the 250 m instrument relative to the 200 m instrument for 13 of the 20 month record (the months with higher transport). The data suggests shear reversal between 200 and 250 m in the western Labani channel, closer to 100 m in the east.
4. The Makassar transport (case B) determined from the Arlindo data is at the higher end of estimates derived from Timor Sea and Indian Ocean studies. While this would favour the case C Throughflow of 6.7 Sv, with zero mean along channel flow at the sea surface this may not be the only explanation. Perhaps we are seeing a Throughflow interannual (ENSO) signal (noting that the JADE Timor Passage values were

obtained during an El Niño period)? Alternatively, might some of the Makassar transport pass back to the Pacific Ocean to the east of Sulawesi? Comparison of the MAK mooring results with: 1996–98 JADE mooring data near Timor (Molcard and Fieux); Lesser Sunda Island shallow pressure gauge array (Janet Sprintall); and data from the Arlindo mooring in Lifamatola Passage (Fig. 1) to be recovered in November 1998, may help resolve this issue.

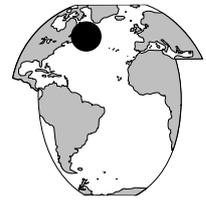
Acknowledgement

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A Western Boundary Current Meter Array in the North Atlantic Near 42°N



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The Newfoundland Basin contains many important parts of the circulation system of the North Atlantic. Along the upper continental slope is the Labrador Current and the western boundary current of the sub-polar gyre of the North Atlantic. Over the deeper continental slope, the Deep Western Boundary Current (DWBC) transports the various components of North Atlantic Deep Water equatorward. These include Labrador Sea Water, Northeast Atlantic Deep Water and Denmark Strait Overflow Water. Near the 4500 m isobath, the North Atlantic Current (NAC) transports warm and salty waters poleward from the subtropical gyre of the North Atlantic.

This region is the transition region between the subtropical and sub-polar gyres of the North Atlantic. The region is bounded on the north and south by the Southeast Newfoundland Ridge and Flemish Cap respectively. Here the continental slope undergoes sharp and rapid changes in direction and steepness, resulting in the formation of meanders, eddies, bifurcations and retroreflections of the major current systems (Clarke et al., 1980; Rossby, 1995).

Because of the complex interactions between the current systems, a control volume study was designed to determine the heat, salt and mass flux through this region. This would serve as the western boundary estimate for the trans-basin section A2 (nominally at 48°N). The control volume study consisted of a series of sections which divided the area from 50°W to Flemish Cap into four closed boxes in which inverse calculations could be performed. These sections would be occupied three times over a two year period.

The most westerly box enclosed the south-east Newfoundland Ridge. Here the Gulf Stream bifurcates and the NAC begins. The northern boundary of this box was chosen to cross the continental slope halfway between the

SE Newfoundland Ridge and the Newfoundland Seamounts. Sea level variability from the Geosat altimeter and hydrographic surveys suggested that this was a region where the NAC was less subject to meanders and eddy shedding. This would allow a moderate current meter array, ACM6, set along this section to capture the NAC transport. This section is also the western end of WHP section A2.

Mooring Array

The mooring array (Fig. 1, page 24) consisted of eight current meter moorings containing a total of 45 Aanderaa current meters. The array was set across the continental slope from the 1500 m isobath (actual depth of 1493 m) to the centre of the Newfoundland Basin at a depth of 4888 m. The inshore limit for the mooring array was determined by the fact that there is active bottom trawling for redfish to depths of 1200 m in this region. This means that the array will not span the core of the Labrador Current usually located on the 1000 m isobath.

The length of the current meter array was 400 km, the separation between moorings varied between 33 and 70 km. The array was deployed in August 1993 and recovered in May 1995 (two moorings whose acoustic releases failed were recovered by dragging in July 1995). The full array provided approximately 640 days (~1.75 years) of data.

Nominal current meter depths of 400 m, 800 m, 1500 m, 2500 m, 3500 m, 4000 m, and 100 m off the bottom were chosen to sample the major water masses and current features. All meters measured speed, direction and temperature. Those at 400 m and 800 m were equipped with pressure and conductivity sensors. Most meters at 1500 m as well as those at 2500 m most likely to be in the core of the NAC were equipped with pressure sensors to document the pull down of the moorings. The mooring lines from 800 m to 400 m on the moorings in the core of the current were faired to reduce drag and subsequent pull down.

In addition to the current meter moorings, six Inverted Echo Sounders were set from the 3300 m isobath to 30 km beyond the most offshore mooring. Three of these IESs were set at current meter mooring locations, two were set between moorings and the sixth was set half a separation distance beyond the final mooring. Only four IESs were recovered and their data is discussed in a PhD thesis

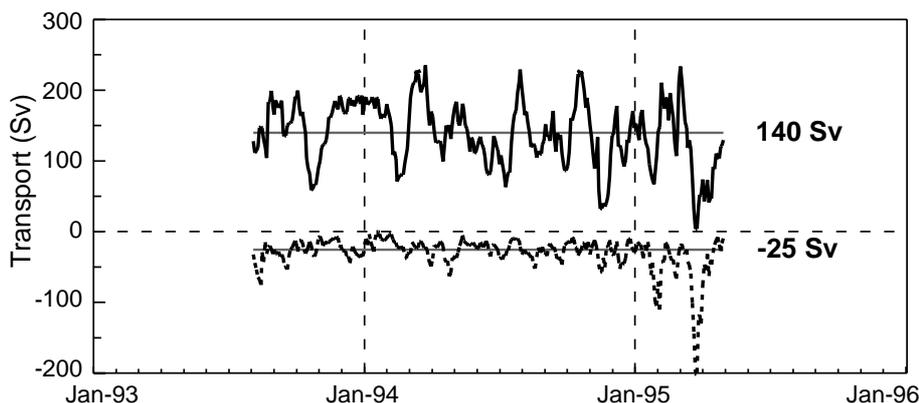


Figure 2. Transport through mooring array ACM6. Solid curve is poleward transport, dashed equatorward transport. The 'mean' value is the median of the individual 2 day estimates of these transports.

from University of Rhode Island (Meinen, 1998).

Data Analysis

Most of the Aanderaa current meters used in this array were the old style tape recording instruments and some of these failed towards the end of the record. Two of the moorings lost their main buoyancy packages during the last three months of their deployment. This loss caused the current meters to sink below the maximum pressure for their pressure sensors resulting in failure of the sensor and flooding of the instruments. The overall data return was 79% for temperature and 76% for velocity.

Due to an oversight, most of the current meters in this array were deployed with a rotor gear ratio that resulted in the rotor counter resetting if the current speed exceeded about 0.3 m/s. This was a serious problem in the upper levels where current speeds of more than 0.7 m/s were frequent. The number of wrap-arounds for these upper current meters was estimated by comparison with the speeds measured at the deeper levels on the mooring and an analysis of the mooring pull-down.

In order to calculate the transport across the section, the temperature and transverse flow observations were low pass filtered and subsampled at 2 day intervals. These were then interpolated, using optimal linear interpolation, onto a regular grid with horizontal grid spacing 20 km and vertical grid spacing 200 dbar.

Total transport estimates require measurements throughout the water column, but no direct measurements were made at levels shallower than 400 dbar. We estimated temperature and flow in the 0 to 400 dbar levels using an indirect method based on climatology. This method assumes that the temperature at 400 dbar can serve as a measure of the location of a particular water column relative to the axis of the NAC. From our climatology of upper ocean sections in this region, we can estimate the geostrophic shear and the temperature structure within the upper 400 dbars as a function of month and 400 dbar temperature. Using these estimates, we can construct temperature and transverse velocity fields in the upper 400 dbars and add these estimates to the gridded fields.

Transports

The NAC is seen in the mean transverse current field as a surface intensified (north-eastward) positive flow with maximum surface speed 0.40 m/s at distance 300 km offshore of the 200 m isobath. The NAC extends to the ocean bottom; the maximum positive flow at the bottom is 0.10 m/s at distance 320 km.

Inshore of the NAC is a region of bottom intensified negative (south-westward) flow. This intensification is associated with the DWBC, marked by a maximum bottom flow of 0.07 m/s at distance 220 km and 4200 dbar pressure.

The transport was decomposed into its poleward and equatorward components. Each elemental grid cell, 20 km \times 200 db \times 2 days, has associated with it a transport.

The positive and negative transports were separately accumulated to provide the time series of synoptic transport, shown in Fig. 2. The median values of positive and negative transport of 140 Sv and 25 Sv respectively are also shown in the figure.

The average transports can also be computed from the means of the 2 day estimates, these have values of 139 Sv and 30 Sv respectively for the poleward and equatorward transports. The difference in the equatorward transport between this estimate and the estimate based on the medians arises from the large equatorward flow observed at the end of the record. The median estimate will be relatively unaffected by this extreme event.

Mean transport can also be defined as the transport associated with the mean flow. This calculation results in much reduced estimates of positive and negative transports due to the smearing of the position of the NAC. Because the flows in the NAC are relatively strong compared to the inshore flows, the transport associated with the negative flow direction in the mean flow field is relatively more affected by this smearing.

Discussion

This preliminary analysis provides a gross estimate of the total volume transport of water poleward and equatorward within the western boundary region around 42°N. 140 Sv is larger than what has normally been considered the transport of the NAC; however, this figure also contains a substantial part of the recirculating flow within the Mann Eddy, a quasi-permanent anticyclonic circulation feature in the centre of the Newfoundland Basin. Reiniger and Clarke (1975) estimated the transport of the North Atlantic Current and Mann eddy across this same section as 121 Sv by referencing a bottle section to current meter data a few 100 m off the bottom.

The equatorward transport of 25 to 30 Sv is also large compared to the estimates of the DWBC transports of 10–15 Sv within the Labrador Sea. This new estimate includes less dense waters than are usually included within the DWBC. In the future, we will make estimates of the DWBC transport in temperature classes.

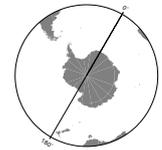
This note is based on material on the web page http://www.mar.dfo-mpo.gc.ca/science/ocean/woce/acm/acm_poster_frame.html which is based on a poster prepared and presented at the WOCE Scientific Conference in May, 1998.

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The Antarctic Circumpolar Wave

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The Antarctic Circumpolar Wave is a recently discovered, large scale atmosphere-ocean phenomenon in the Southern Ocean. It is perhaps the newest member of the growing family of identified interactive processes that affect the climate of the globe (such as ENSO and the North Atlantic Oscillation). Most of the observational details have been described by White and Peterson (1996) and Jacobs and Mitchell (1996).

The observed ACW is perhaps most prominent as a zonal wave number two pattern in sea surface temperature (SST) anomalies (i.e. with the seasonal average removed) of typically plus or minus one degree in the Southern Ocean, moving slowly eastward (speed ~ 9 cm/sec or 40° longitude per year), taking about 9 years to circle the pole. This is comparable with the mean speed of the Antarctic Circumpolar Current (ACC). At a fixed location, a single cycle takes approximately four years. Together with the SST, there is a corresponding variation in the sea surface height (in phase), the northward sea ice extent (in antiphase), and the atmospheric sea level pressure and the wind stress curl, which both lead SST by about 1 year, or $1/4$ cycle. The sea ice extent is apparently related to the ease of access by shipping to the Antarctic coast.

All of these observations are of course somewhat “noisy”, and the reported results depend on filtered data. But the underlying phenomenon seems to be quite robust. Fig. 1 (page 23) shows composite satellite pictures of SST anomalies for the southern hemisphere for two weeks, one in January 1997 and the other in January 1998. Alternating warm and cold “blobs” can be clearly identified as the ACW in the Southern Ocean, moving eastward at the expected speed.

The reason as to why such a large and apparently significant phenomenon as the ACW could have escaped detection until now are fairly clear – there have been so few observations in the Southern Ocean that one needs appropriate satellite technology to see it. Suitable observations of sea surface temperature and height (via altimeter) only go back about 12 years – barely enough to resolve a 9-year period. It is possible that the structure was different at earlier times. However, there are indications of 4-year (approx.) periodicity in oceanic carbon isotope ratios in longer records, so that the wave number two structure may be quite persistent.

The extent to which the ACW is linked to other phenomena such as ENSO or the Indian Ocean dipole (Nicholls, 1989) has yet to be determined. However, there is evidence that it has greater impact on the temperature and rainfall of New Zealand than that of ENSO (White and Cherry, 1998), and one may expect similar results for southern Africa, southern Australia and southern South

America. Warmer water to the south should result in warmer and wetter southerly winds on land, and conversely for colder water.

What causes the ACW? One possibility that has arisen from some model studies (Weisse et al., 1998) is that it is due to a low frequency SST response to high frequency atmospheric forcing, as described by Saravanan and McWilliams, 1998), and the idea dates back to Hasselmann (1976). Essentially, in this process SSTs embark on a low-frequency random walk forced by high-frequency synoptic disturbances in the atmosphere. If this is the case, the ACW would be essentially unpredictable, since it would be necessary to first predict the weather. It is likely that this process works to some extent, but that its amplitude is limited.

Fortunately (from the viewpoint of predictability), there is another possibility, namely that the ACW is due to a positive feedback between the ocean and atmosphere. A stability analysis of a two-layer model in which the upper layer is a barotropic atmosphere and the lower is an upper layer of the ocean, in which the two interact with wind stresses and surface heat fluxes, produces unstable solutions (Baines and Cai, 1998). These disturbances have maximum growth rates at long wavelengths (such as zonal wave numbers two and three) in which barotropic Rossby waves in the atmosphere are almost stationary, and the wind stress forcing of the ocean can resonate with the heat flux forcing of the atmosphere. The most rapidly growing disturbances have a similar structure to the observations described above.

Various forms of an ACW have also been reported in the current generation of coupled atmosphere-ocean models. Here they all seem to have zonal wavenumber three, with no conspicuous propagating signal in the atmosphere. In the CSIRO model, in particular, the oceanic part of the ACW is very deep and contains considerable heat storage, while the atmospheric signal is a stationary wave number three, with the regions over the three oceans oscillating in phase with the temperature of the ocean below (Cai, Baines and Gordon, 1998). This atmospheric signal has very similar character to that described by Mo and White (1985). The mechanism described above can explain this behaviour if one first assumes that the atmospheric response has the structure of a stationary oscillating pattern. It is possible that, in the real world, both stationary and propagating atmospheric signals interact with the moving ocean anomalies.

Coupled models are allowed to have their own peculiarities, and the challenge is to use them to learn about the interacting atmosphere-ocean (and ice) system. The Antarctic Circumpolar Wave is far from being a closed book.

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WOCE-AIMS Tracer Workshop, Second Announcement

As announced previously (see WOCE Newsletter No. 31, June 1998, p.17), the WOCE-AIMS Tracer Workshop will be held at the University of Bremen, Bremen, Germany, 22–26 February 1999. The objectives are:

- (1) Further the oceanographic interpretation of the WOCE tracer data sets in continued co-operation with the physical oceanographic community;
- (2) Encourage the synthesis, compilation and dissemination of tracer data sets;
- (3) Establish mechanisms for the production of data products (gridded and derived) for the use of a broader community;
- (4) Identify strategies to facilitate assimilation and incorporation of tracer data into models;
- (5) Initiate collaboration and interaction between tracer geochemists and physical and chemical oceanographers and modellers.

The workshop is aimed at setting a foundation for the further evaluation and use of the WOCE tracer data set in order to optimise scientific output over the WOCE-AIMS period, rather than on presenting and discussing individual tracer projects. Central questions will be in which way and to which extent tracer data are useful to yield information on large-scale circulation and transport (and their variability), (1) in a hydrographic context (including parameters such as nutrients and oxygen), and (2) in models, and furthermore how they can help to constrain and improve models. Interaction and initiation of scientific collaboration between workshop participants will be a particular concern. We hope that the entire community involved in WOCE work related to tracers including physical and chemical oceanographers and modellers will be represented at the workshop. Contributions from colleagues critical to the tracer approach are welcome.

A planning group, consisting of Wolfgang Roether, Bremen, Germany (co-chair; wroether@physik.uni-bremen.de); Scott Doney, Boulder, USA (co-chair; doney@ncar.ucar.edu); William J. Jenkins, Southampton, UK; Yukata Watanabe, Ibaraki, Japan; William M. Smethie, Palisades, USA; Claus Böning, Bremerhaven, Germany;

and Matthew England, Sydney, Australia, has agreed on a format of the workshop as follows:

The stage for discussions in keeping with the workshop objectives will be set by invited talks addressing

- (1) The present status of tracer oceanography,
- (2) Open issues in physical and chemical oceanography and role of tracers,
- (3) Model-data synthesis: methodology and directions, and
- (4) Tracer data dissemination and products.

Working groups organised in correspondence will follow. Posters are invited addressing

- (1) research group overviews on available datasets and ongoing tracer data evaluation projects,
- (2) methodical developments in tracer data evaluation, and
- (3) specific research projects related to the subject of the workshop.

WOCE IPO has requested a written report from the workshop. In preparation for this and to structure the discussions, speakers and working group chair persons will be asked to prepare extended abstracts for distribution to participants prior to the workshop.

More detailed information on the programme and on local arrangements (travel, housing, etc.) can be found in <http://pacific.physik.uni-bremen.de/workshop/index.html>; an electronic registration form is included. Participants are asked to register before 10 December. Bremen can be reached by air from various major European airports; note that in February the weather may be rather miserable and cool.

Participants are asked to provide their own travel funds as far as possible. We can help out a limited number of participants with partial support (at most air ticket + hotel). We have to receive requests for support by the registration deadline. Further support may be available through national WOCE channels. Attendance at the workshop will be limited in principle, but we hope to be able to accommodate everybody interested in attending.

Towards the Definitive Space Gravity Mission

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For many years, a large part of the geophysics community, including oceanography, has been lobbying for a space gravity mission which would provide a precise description of the Earth's gravity field and geoid independent of the quasi-geoid information provided by altimetry. Proposals such as GRM, Gravsat, GAMES, Gradio and Aristoteles have come and gone, killed off either by budget reductions or international politics.

It was recognised twenty years ago that space gravity would be an essential complement of precise altimetry. In the original TOPEX 'grey book' proposal (TOPEX, 1981), it was assumed that Gravsat would fly at more or less the same time as TOPEX, thereby providing gravity field models which would improve significantly the altimeter orbit errors achievable at that time. The 'grey book' also pointed to the fact that mean sea surface (MSS) height minus geoid height would supply oceanographers with the ocean dynamic topography, freeing them from having to make assumptions on 'levels of no motion' in hydrography. Wunsch and Gaposchkin (1980) had shown how a formalism could be constructed for including estimated geoid errors, and errors in MSS and hydrographic fields, in computations of dynamic topography.

Interestingly, now that TOPEX/POSEIDON has achieved excellent orbit error reductions, via a programme of gradual but sustained gravity field improvement by conventional methods together with the development of advanced forms of tracking (DORIS, GPS), the oceanographic justification for a gravity mission is stronger than ever. There are three main reasons:

1. To take advantage of the centimetric accuracy MSS fields now available by provision of centimetric geoids. One could have argued that the poorer MSS fields available some years ago would not have justified expenditure on a gravity mission in this way.
2. To further reduce orbit errors of the lower flying altimeter satellites (ERS-1/2, Envisat, Geosat Follow On) to the level achieved for TOPEX/POSEIDON. Residual gravity model errors are still major factors in orbit determination for the lower flying missions, even given near-global tracking.
3. To take advantage of the synergy of scientific knowledge which will be obtained by such missions now that, twenty years later, they are cheaper and technically more feasible. For example, knowledge of processes in the solid Earth such as Post-Glacial Rebound obtained from a mission such as the European Space Agency's GOCE (Gravity field and steady-state Ocean Circulation Explorer, see ESA, 1996) will benefit oceanographers and climatologists interested in global sea level change. In addition, knowledge of the temporal dependence of the gravity

field obtained from the US-German GRACE (Gravity Recovery And Climate Experiment, see NASA, 1996) will provide data on movements of mass in the ocean, in the ice caps and on land which will have a wide range of application across geophysics.

The reader may know that three space gravity missions, the German CHAMP (CHALLENGING Mini-satellite Payload, see Reigber et al., 1996) mission as well as GRACE and GOCE, are currently proposed for launch within the next 5 years. This much improved situation may stimulate the reader, who is probably an oceanographer and also a taxpayer, to ask why three are needed when oceanography seems to have been rubbing along fairly well for so long without one. For the detailed scientific arguments, we refer the reader to ESA (1996), NRC (1997) and Dickey et al. (1998); we believe the arguments will be convincing. We shall concentrate in this note on the topic of providing a precise, high resolution gravity field or geoid. (For discussion of monitoring temporal changes in gravity with long duration missions such as CHAMP and GRACE, see NRC, 1997).

The purpose in writing this note is to emphasise the unique contribution to the recovery of the gravity field and geoid from GOCE, as well as to document the complementarity with GRACE and CHAMP. In particular, the three missions offer major differences in the recovered resolution and accuracy of the gravity field spectrum. However, we do want all three. In addition, to be realistic, we know that there will inevitably be a learning curve between missions in space gravity (as there was in altimetry). No oceanographer would have settled for the Seasat data set if he had known TOPEX/POSEIDON was a few years away. Space gravity over the next few years will culminate in what will be, in our opinion, the definitive GOCE mission.

The different missions

There are three missions planned: CHAMP, GRACE and GOCE.

CHAMP is a low cost, multi-payload, small satellite mission, providing a gravity field intermediate between our present knowledge and oceanographic requirements (which are summarised below). Launch is planned for late 1999 with the mission lasting 4–5 years. Altitude will be approximately 470 km initially, reducing with air drag to 300 km.

GRACE is a more advanced mission, especially aimed at monitoring the time variations of the gravity field at long wavelengths (i.e. 500 km and longer). It consists of two CHAMP-like satellites about 250 km apart connected by a satellite-to-satellite (SST) microwave link. The altitude

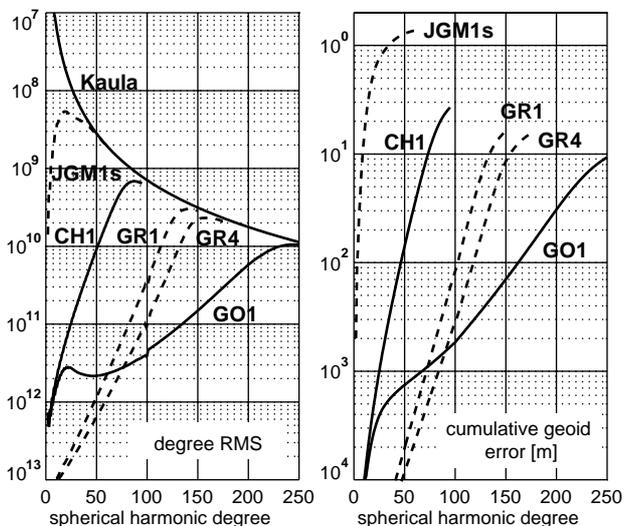


Figure 1. Spectral error results of the baseline missions: dimensionless degree RMS curves (left) and cumulative geoid errors, or commission errors (right). From ESA (1998).

will initially be around 470 km decaying towards 320 km at the end of the mission. Launch is planned for mid-2001 with the mission lasting 3–5 years.

GOCE is a high resolution gravity field mission and will open a completely new range of spatial scales (order of 100 km) of the gravity field spectrum to the research community. It consists of a single satellite of high mass-to-area ratio, with either a ‘room temperature’ (capacitive) or superconducting (inductive) three-axes gradiometer. Launch is planned for 2003 with the mission lasting 8 months. Altitude will be kept at about 250 km.

All three missions will have a near-polar orbit. Each will use GPS (or GPS/GLONASS for GOCE) high-low SST, providing the longer wavelength components of the gravity field. The two GRACE satellites also employ SST in low-low mode to recover the shorter wavelength components. GOCE employs gradiometry to provide the latter. CHAMP and GRACE will use accelerometers to measure the non-conservative forces (drag) operating on the spacecraft; GOCE will use the ‘common-mode’ capability of the gradiometer to measure the remaining non-conservative forces after its Drag Free Control has compensated for most of them.

It has proved extremely difficult to make simulations which might provide meaningful comparisons between the different missions with regard to achievable gravity field recovery. For example, not only are the different satellites planned to fly at different altitudes and at different points in the solar cycle, but the altitudes of CHAMP and GRACE will reduce during their missions, with, in principle, greater precision being obtained in the later stages. Therefore, one has to assume that the satellites will have lifetimes as anticipated. In addition, GRACE and GOCE will be using technologies which have never been used in space before, and the estimation of instrument performance (microwave SST link for GRACE, gradiometer for GOCE) is critical to

the simulations: should one assume target errors for these technologies or be conservative?

In spite of these reservations, a set of simulations has been carried out recently in order to identify the strengths and weaknesses of the different proposals (Balmino et al., 1998). First, a set of ‘normalised mission concepts’ was constructed. These simulate idealised missions of the same duration (30 days) and precise polar orbit, and with nominal accuracies for GPS differential positioning, GRACE-like SST, and GOCE-like capacitive or inductive gradiometry. The results confirm what has been known for some time, that GRACE-like SST is superior to GOCE-like gradiometry in the lower harmonics below degree and order typically 50–60 (equivalent half-wavelength of approximately 400 km), making a GRACE-like mission optimal for studying time dependent gravity errors at long to moderate wavelengths. Gradiometry, on the other hand, is superior for studying high spatial resolution features as small as 100 km half-wavelength, and especially those which are not time dependent.

Our studies then progressed to perform specific CHAMP, GRACE and GOCE simulations using a range of instrument performance characteristics, non-polar inclination orbits and altitudes. The results are summarised in Fig. 1 wherein JGM1s refers to the estimated accuracy of the currently available JGM1s model based purely on orbit information; CH1 refers to CHAMP; GR1 and GR4 refer to GRACE with a 400 and 320 km altitude respectively; and GO1 refers to GOCE with the non-superconducting gradiometry. (See Balmino et al., 1998 for full details of parameter values adopted in the simulations.)

From Fig. 1, it is again obvious that GRACE is superior for the low degrees, say up to 50. This is not strictly an intrinsic feature of SST low-low, but is rather a result of the extraordinarily high assumed system performance advertised by the mission.

GOCE, on the other hand, outperforms all other missions in the higher degrees up to degree 250, with the error curves for GRACE and GOCE crossing between degrees 60 and 80, depending of course on the specific mission parameters. A lower orbit, or better measurement accuracy, or scaling of the mission duration would push the GRACE curve downwards. However, since the curves are steep, the cross-over point would shift to the right hand side by a relatively small amount.

The oceanographic requirements

It is clear that each of these missions will result in major gains in knowledge of the gravity field and geoid, but what requirements do oceanographers really have?

Most oceanographers will know that at present an altimetric MSS is distinguishable from the best model of the geoid up to degree 15 or so (or wavelengths of approximately 2000 km); at shorter scales the errors in the geoid models render such subtractions imprecise (see ESA, 1996 for a review). Recent studies, which do not differ qualitatively from others performed over the last 20 years,

have expressed geoid measurement error requirements as a spectrum of magnitude approximately 2 cm averaged over wavelengths of 100 km; 0.2 cm over 200 km wavelengths; through to less than 0.1 cm at 1000 km wavelength (the ‘basin scale’). The short 100 km scale is often referred to, perhaps inappropriately, as the ‘mesoscale’; it is intended to represent essentially the Rossby radius. Fig. 2 indicates schematically the signals in the dynamic topography which we wish to identify via studies of MSS minus geoid. These vary from short wavelength features such as through-flow currents, coastal currents and deep ocean fronts to the large scale ocean gyres.

For example, Wunsch (quoted in ESA, 1996) has shown that an error of 1 cm in the geoid height difference (or altimetric MSS height difference) across the North Atlantic at 30°N corresponds to a volume transport of 7 Sv (if interpreted in a barotropic sense) and approximately 10^{14} W in meridional heat transport. These are large numbers but measurements to this accuracy would represent significant improvements compared to present uncertainties. It is clear that one has to do better than 1 cm at these scales, hence the 0.1 cm requirement at 1000 km wavelength.

Developments for merging such information at this scale, which will be obtained by both GRACE and GOCE, into ocean models have recently been discussed by Ganachaud et al. (1997), Wunsch and Stammer (1998) and Le Grand and Minster (1998). Fig. 3 illustrates an example of such work using a coarse (4.5°) inverse model. The figure shows transport uncertainties (Sv) estimated by the model using the EGM96 and GOCE error budgets (dark and light grey respectively). The left hand panels show uncertainties associated with zonally integrated transports across 24°N, 36°N, and 48°N, while the right hand panels show the uncertainties associated with meridional transports across 36°N in the region of the Gulf Stream between 75°W and 72.5°W. From top to bottom, the panels show transport uncertainties integrated from the surface to the ocean bottom, from the surface to 100 m depth, from the surface to 1000 m depth, and from 3000 m to 4000 m depth. The reduction in uncertainty obtained at 48°N section when the GOCE error budget is used is shown in percent. Uncertainties corresponding to EGM96 and GOCE error budgets are not significantly different for surface to bottom transports (top panels), and for deep ocean transports (bottom panels). However, they are significantly different for transports in the upper ocean (middle panels), especially for the 48°N section. In this section, the

uncertainty in transports in the upper 100 m of the ocean is reduced by 26% when the GOCE error budget is used. The corresponding volume transport uncertainty reduction is about 0.2 Sv, which translates into a heat transport uncertainty reduction of about 10^{13} W.

This present impact estimate, which is based on available data and error budgets only, is a conservative one for several reasons:

- (i) Uncertainties in Ekman transports will be reduced in the near future using new scatterometer data. Neglecting Ekman transport uncertainties doubles the impact of GOCE on volume flux uncertainty reduction at 48°N (52 instead of 26% reduction).
- (ii) Ganachaud et al. (1997) showed that the North Atlantic is the ocean basin where gravity missions will have the smallest impact.
- (iii) The calculations use estimates of mean dynamic topography averaged over 4.5°, and the present study therefore underestimates the impact of GOCE on the determination of transports along sharp fronts like the Gulf Stream.

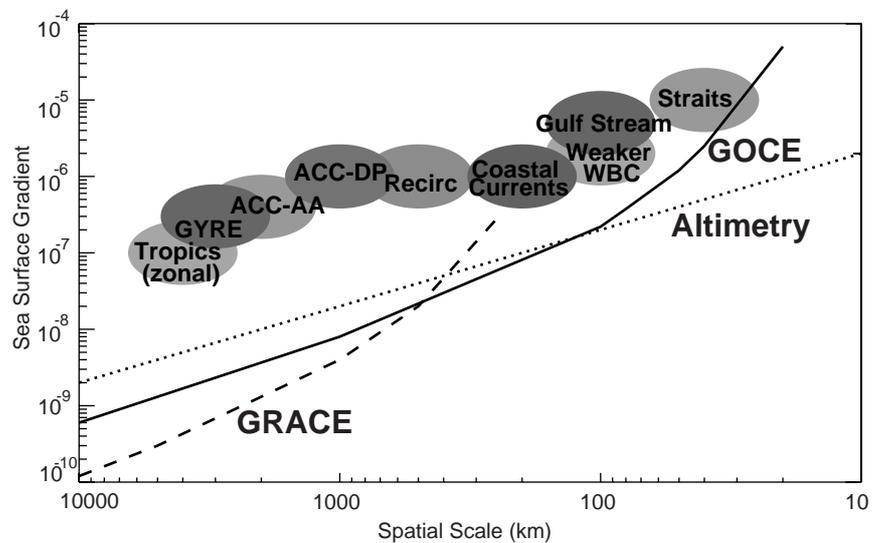


Figure 2. Highly schematic illustration of sea surface gradients (relative to the geoid) of several components of the ocean topography compared to MSS slope accuracy from altimetry (dotted line) and to geoid slope accuracy from space gravity missions such as GOCE (solid line) and GRACE (dashed line). ‘Gulf Stream’ represents the stronger deep ocean fronts including those of the Gulf Stream itself and of, for example, the Antarctic Circumpolar Current. ‘Recirc’ represents the Gulf Stream recirculation. ‘Weaker WBC’ represents the weaker Western Boundary Currents (e.g. Brazil Current) with spatial scales of order 100 km and gradients of order 10^{-6} . ‘ACC-DP’ and ‘ACC-AA’ represent a major current such as the ACC at Drake Passage or at the wider African and Australian choke points respectively. ‘GYRE’ represents a typical 1 m ocean gyre over 3000 km scale. ‘Coastal Currents’ represents the myriad of coastal currents, flows through longer straits and meridional equatorial signals with space scales of order 100 km and gradients of 10^{-6} . ‘Straits’ represents flows through short straits which are at the limit of spatial resolution. GOCE will be needed to resolve all these signals. Note that at very long wavelengths, where GRACE accuracy is superior to that of GOCE, remaining altimeter orbit and other systematic uncertainties are still significant.

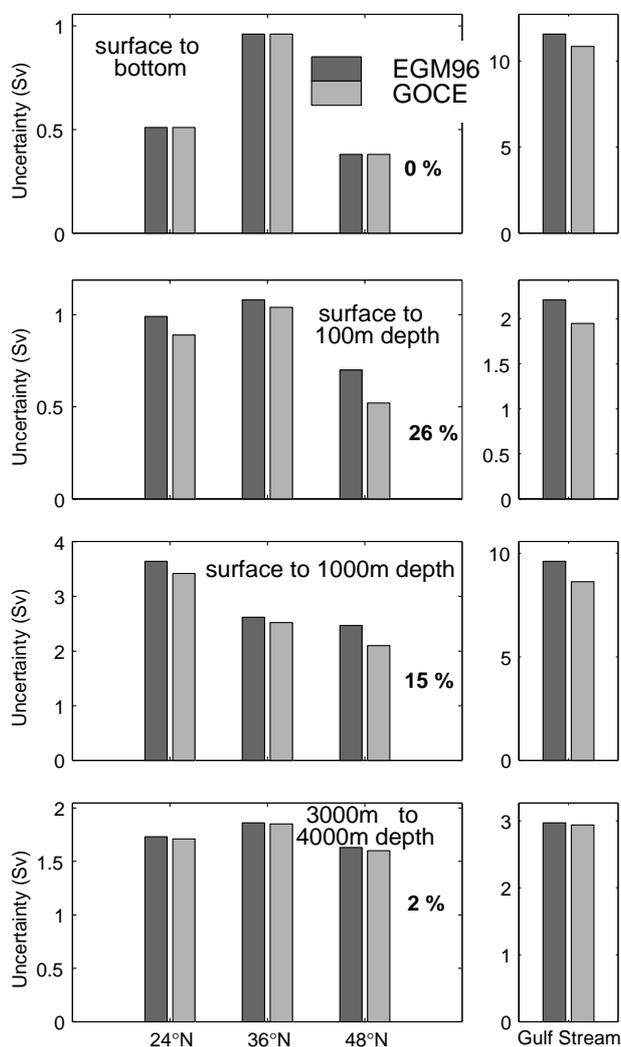


Figure 3. Estimates of reductions in transport uncertainties with the use of a gravity mission such as GOCE (see text for explanation).

(iv) Because the present study is a steady state one, it must account for the presence of noise in the density field caused by natural variability. New assimilation techniques (Hasselmann and Giering, 1997; Wunsch and Stammer, 1998), combined with precise estimates of the instantaneous density field from either better sampling of modelling, will circumvent this problem and allow the MSS - geoid information to reach deeper levels of the ocean.

Therefore, in summary, is it really necessary to resolve the small spatial scales with a gravity mission? Is their resolution essential to our understanding of the ocean circulation, and climate? Will not such small scale features become evident in models once the large scale flows are adjusted? These are questions which modellers involved in assimilating potential gravity fields into ocean models are endeavouring to answer. What is clear is that if one simply high-pass filters a dynamic topography from a long run of

a numerical model (e.g. Semtner-Chervin) with a filter which preserves signals of degree 80 or more (i.e. the 'GOCE-only-accessible' part of the spectrum of Fig. 2), then many important short spatial scale features are evident (frontal signatures of the major currents; definition of narrower and smaller boundary currents; ACC jet banding; zonal equatorial signals) (Fig. 4, page 24), and such models cannot, of course, be claimed to be a complete representation of the ocean.

It is our belief that the correct scientific approach is to measure the shorter wavelength components of the gravity field spectrum if one has the means to do so (i.e. by means of a mission such as GOCE), rather than rely on their simulation via model constraints. We would appreciate receiving your views on this question.

It is also our belief that GRACE and GOCE together would provide an outstanding data set, covering all parts of the gravity field spectrum with unprecedented accuracy. This combination would really be the 'definitive mission'. We hope that GRACE, GOCE and the pioneering CHAMP mission will all receive the support of the oceanographic community.

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The WOCE North Atlantic Workshop, 23–27 August 1999 in Kiel, Germany

The purpose of the workshop is the exchange of scientific results on North Atlantic water mass transformation, circulation and meridional overturning that have been obtained by various observational means, by numerical modelling and by analysis of the climate record, with the main emphasis on the WOCE period. Central objective is progress toward a synthesis regarding oceanic flux divergences and the implications for the atmosphere. A specific set of workshop foci and an agenda are being formulated by the organising committee.

Format

The workshop will consist of invited lectures, poster group presentations and working group discussions centred on key scientific topics, and aiming at enhancing interactions between research groups to accomplish the overall synthesis objectives. Detailed structure and a list of workshop themes will be announced in a later WOCE Newsletter issue.

Participation

To accomplish an interactive and stimulating workshop atmosphere it is expected that participants are actively involved in research relevant to North Atlantic WOCE objectives, and prepared to participate constructively in working group discussions. Potential participants should contact Chair of Organising Committee or Session Chairpersons (TBA in later Newsletter) about their interest and proposed input into the workshop, i.e. by providing data analysis products, model results etc. as basis for workshop discussions.

Scientific Organising Committee

C. Böning, IfM Kiel, Germany;
H. Bryden, Southampton Oceanography Centre, UK;

R. Molinari, NOAA/AOML, Miami, USA;
G. Reverdin, LEGOS, Toulouse, France;
P. Schlosser, Lamont Doherty Earth Observatory, Palisades, USA;
F. Schott, IfM Kiel (Chair);
C. Wunsch, MIT, Cambridge, USA.

Kiel and the meeting venue

The workshop will be held at the Institute für Meereskunde with the main sessions taking place in the neighbouring Art Gallery.

Registration Fee and participation support

A modest registration fee will have to be raised to cover some of the costs of the event. Some funds will be available from the WOCE IPO to support individuals travel and accommodation costs.

Accommodation

Blocks of rooms have been reserved in various Kiel hotels. Information can be soon obtained from a Website that is presently being established. Participants will then be expected to make their own arrangements.

Local organisation

Prof. F. Schott e-mail: fschott@ifm.uni-kiel.de.
Secretariat: Mrs K. Maass, e-mail: kmaass@ifm.uni-kiel.de;
Mrs S. Komander, e-mail: skomander@ifm.uni-kiel.de,
Tel./Fax 49 (0)431 597 3821.

A Website on Workshop logistics and further organisational details is presently under construction and will also be accessible from <http://www.soc.soton.ac.uk/OTHERS/woceipo/ipo.html>

Report on Indian Ocean Workshop

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The WOCE Indian Ocean Workshop was held at the Hampton Inn, Carondelet St., New Orleans during 22–25 September 1998, with financial support for the logistics provided by the US WOCE Office. Some 56 people attended from the US, Australia, France, Germany, India, Japan, and the UK. Attendees were reminded of the importance of air-sea interaction in the northern Indian Ocean by the looming threat of Hurricane Georges, which passed over the Florida Keys during the workshop and dropped large quantities of rain on the coast of Mississippi immediately after the workshop finished. Over one million people were evacuated from the coastal regions of Mississippi and Louisiana during this process!

The meeting was arranged to include plenary talks, discussion periods, and free time for attendees to get together and talk about science. Plenary talks were given by Greg Johnson (Meridional overturning), John Toole (Transports of energy, freshwater, and constituents), Fritz Schott (The response of the north Indian Ocean to the monsoons), Don Olson (Non-monsoonal variability), and Jochem Marotzke (Development and testing of Indian Ocean models). About 30 attendees brought posters to the meeting; these were on display throughout and generated much discussion during breaks and in the afternoons. It is suggested that attendees make their posters available electronically via the IPO site on the web.

Two large discussion groups coalesced during the meeting. The first considered the meridional circulation of the region, the second concentrated on variability within the Arabian Sea/Bay of Bengal and its connection to the Indonesian Throughflow. Additionally, many smaller groups of investigators met over lunch and dinner to discuss new proposals, analyses, syntheses and the like. It was encouraging to see a large percentage of younger researchers among the attendees, and the organisers hope that they will take the lead in future work in the region. Overall, the amount of discussion between attendees was good, and it is assumed that the real products of the workshop will be the analyses, syntheses and new scientific interpretations that result from them.

While WOCE research in the Indian Ocean has undoubtedly given us a large incremental increase in our knowledge base of the region, there was considerable discussion of continuing gaps in our knowledge. The attendees made certain recommendations for future work, mostly concerned with establishing levels of variability. These included: Continued monitoring of the transport and T/S structure of the Indonesian Throughflow; The establishment of a permanent array across the equator south of Sri Lanka to monitor changes in the monsoon current; Continued occupation (and preferable augmentation) of VOS lines, with salinity measurements if possible; and Continued deployment of PALACE floats and drifters and long-term in situ sea level, SST and satellite measurements.

Participants were also concerned about certain elements in the original observational plan that, for one reason or another, had not been accomplished. They urged that interested investigators be encouraged (and funded) to carry them out, and thus complete the WOCE programme. These are: CTD sections across the South Java Current; Array of moored current meters to measure the Indonesian Throughflow into the Indian Ocean; Array of moored current meters to measure the net flow through the Mozambique Channel; Completion of WHP section I7 south to Antarctica; and Occupation (with the full set of WHP measurements) of the central portion of section I5.

During the workshop, representatives of the WOCE atlas committee met to agree on common formats, colours, etc., and made a presentation to a plenary session of the workshop. Attendees accepted the need for electronic atlases, but strongly endorsed the idea of paper atlases at a scale sufficiently large that they can be used to examine small-scale structure. This demands maps similar in size to those produced following the Scorpio Expedition or during IGY – approximately poster sized. This would allow even the long lines in the Pacific to be produced at vertical exaggerations of 1:1250 in the upper 1000 metres and 1:500 throughout the rest of the water column. Such large maps could perhaps be produced as unbound charts to reduce costs.

The full report on the meeting is presently being written and will be made available as soon as possible.

The on-line WOCE Bibliography

Found on the Web at http://oceanic.cms.udel.edu/diu_cd/diu/biblio*

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The International WOCE bibliography was started in October 1994 as a result of an initiative of the current Director of the International Project Office (IPO) which was approved by the WOCE Scientific Steering Group. Its purpose is to provide a convenient reference source for WOCE Investigators as well as a record of the achievements of the project.

The bibliography is a collaborative project between the IPO, the Data Information Unit (at the University of Delaware) and the UK National Oceanographic Library at the Southampton Oceanography Centre. As of 1 September 1998 it contained 3045 references.

The growth of the bibliography

Back in the late 80s and for the first year or two of the 90s most of the references loaded into the bibliography database were reports of WOCE planning committees, both international and national, of conferences and working groups, many concerned with TOPEX/POSEIDON and

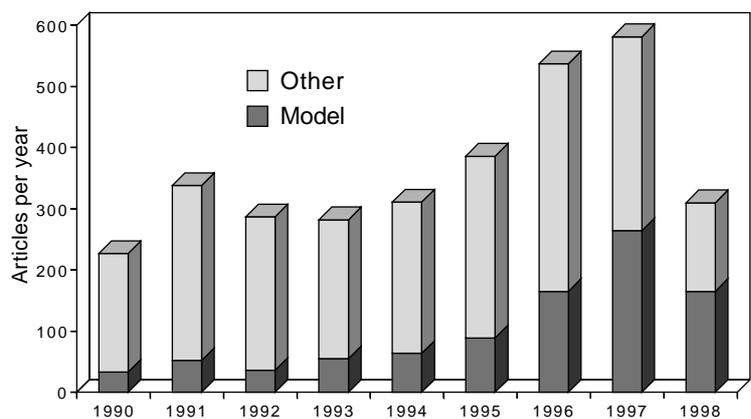


Figure 2. The modelling content of the WOCE bibliography.

* A version of the bibliography is available on the CD-ROM set issued at the WOCE Conference, Halifax, May 1998 (DIU disk). This will be updated as new versions of the CD-ROMs are issued.

ERS-1 satellites.

Gradually reports of scientific results began to emerge, sometimes in the International WOCE Newsletter, and then in the refereed journals. All articles in the journals which were concerned with large scale ocean circulation or the large scale forcing of the ocean were and continue to be included, whether the research was or is sponsored by WOCE or not. Publications from the TOGA and TOGA-COARE projects for example are included. The rationale for this is that such research is almost certainly of importance to the WOCE community and that exclusivity is unhelpful to good science.

Nevertheless it was deemed important to distinguish the scientific output resulting from the project. Based on a knowledge of who was working in WOCE, on lists of publications supplied by WOCE National Committees, on printed acknowledgements of WOCE support and on articles using WOCE data, some fraction of the refereed articles have been identified as originating in and sponsored by WOCE.

The accompanying Fig. 1 (page 22) shows the growth of the bibliography in three categories, namely non-refereed articles, refereed articles due to WOCE and refereed articles originating outside WOCE. At the time of writing, October 1998, the totals in these categories are in round terms 1400, 700 and 900 respectively, totalling 3000 in all. The growth of the scientific output is evident as is the increasing number of refereed articles. Few will deny that the intellectual content of this collection is the most important component of the legacy of WOCE.

Subjects (keywords)

As with all modern library collections each article is associated with a list of subjects (also known as keywords). This information is provided by the UK National Oceanographic Library at the Southampton Oceanography Centre (formerly at IOS). And of course search software is needed too (courtesy of the Data Information Unit, Delaware). Recently the DIU has loaded the bibliography onto a new server and (once you can get through on the WWW) searching is very rapid and lists of a particular subject can be generated with ease.

Models

Apart from the categories described in the previous section one of the largest lists is generated by the subject word 'model'. This captures articles on ocean, coupled and climate modelling and totals 900 items (refereed and unrefereed articles). The size of this field is astonishing and its contribution to the growth of the bibliography is seen in Fig. 2. In the last two years almost one half of the articles acquired are related to modelling. A new and growing topic in this field is data assimilation, and so far 80 articles have been so labelled.

Oceans (as of 1 September 1998)

Another set of large groupings is found by searching on each ocean.

Antarctic (Southern Ocean)	243 items
Atlantic Ocean	865
Indian Ocean	238
Pacific Ocean	651

Each ocean can also be searched by region North, South, tropical and further sub-divided East and West. The above totals about 2/3 of the bibliography but sometimes we don't manage to label every item and some quite correctly get more than one ocean keyword.

Observational techniques

Here we can construct a long list but we hope the following will cause you to reach for your Web browser.

Satellite (altimetry, SST, winds)	243
ADCP	91
Drifters	87
Floats	32
Current Meters	31
CFCs	118
Water masses	184
Transport	158
WOCE Section (e.g. SR01)	145
Mixing	66
etc., etc., etc....	

And a matter for pride for the IPO staff is the entry under WOCE International Newsletter, a total of 334 articles.

Country

Here are the contributions from the top 8:

Australia	107
Canada	123
France	177
Germany	338
Japan	133
Russia	108
UK	700
USA	1387

The WOCE bibliography is a list of articles printed or published in English so one half of the countries above will have additional articles in their own language and so will be under-represented to varying degrees. In addition the UK contribution is swollen by closeness to source and the blanket WOCE attribution to publications for the years 1990-97.

Acknowledgements

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Note on Copyright

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<http://www.soc.soton.ac.uk/OTHERS/woceipo/ipo.html>

We hope that colleagues will see this Newsletter as a means of reporting work in progress related to the Goals of WOCE as described in the Scientific Plan.

The editor will be pleased to send copies of the Newsletter to institutes and research scientists with an interest in WOCE or related research.